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PIEDMONTITE IN ARIZONA

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Piedmontite the rare manganese member of the epidote group was first described from Arizona by Dr. C. Lausen in 1927.[1] This occurrence was in andesitic rock on the southeast slope of Pat Hills in the Sulphur Spring Valley about 25 kilometers northeast of Pearce, Arizona. It was later discovered by Dr. W. H. Brown in the Tucson Mountains immediately west of Tucson in rhyolitic rocks associated with sandstone or quartzite,[2] and still later, in 1934, by the present writer in the Santa Rita mountains about 50 kilometers south of Tucson. This paper contains a brief description of the Santa Rita occurrence as well as a more detailed account of the striking features displayed in the Tucson Mountain deposits. The latter is given greater prominence on account of its greater distribution as well as its more remarkable paragenetic interest.

Piedmontite seems to have been known to Cronstedt in 1758 from its earliest locality, Piedmont, Italy. He refers to it as "röd Magnesia." This work appears to have been forgotten until 1790, when Napione refers to it as "Manganèse rouge" and publishes analyses. On the basis of these analyses, Haüy in 1901 gave it the name of "Manganèse oxydé violet silicifère." In 1803, Cordier from the results of new analyses renamed it "Epidote Manganésifère," a name which was adopted by Haüy. Thus at this early date the true nature of the mineral was appreciated. Later the mineral became known as "Piedmontischer Braunstein" and finally as Piedmontite.[3]

A specimen of piedmontite from San Marcel, Piedmont, Italy, in the writer's possession appears as a schistose or gneissic rock consisting chiefly of quartz and piedmontite, the latter being oriented according to the schistosity of the rock. It is also found in mica schists on the Island of Groix and in the chlorite and glaucophane schists of Japan.[3] Rosenbusch claims it as a petrographic mineral as shown by his use of the terms, Piemontitschiefer, Piemontitphillite and Piedmontitquarzit.[4] On the American Continent, Williams described it from the rhyolites of South Mountain, Pennsylvania,[5] and Harworth from the quartz porphyries of Annapolis, Missouri.[6] Rogers also found it in thin sections of quartz porphyry from a boulder in gravels at Pacific Beach, San Diego Co., California.[7] More recently Mayo describes it as occurring in a "piedmontite sericite schist of volcanic origin" from near Shadow Lake, Madera Co., California, in crystals .2 to .3 mm. in length. It was also found in a second locality in the same county replacing biotite in a similar rock.[8] Material from this locality was further investigated as to optical properties of piedmontite and their relation to manganese contents, by A. M. Short.[9] It would seem, then, that its appearance in rhyolitic rocks or closely related volcanics is perhaps its most characteristic mode of occurrence in this country.

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[1] Lausen, C., Piedmontite from the Sulphur Spring Valley, Ariz.: *Am. Mineral.*, vol. **12**, pp. 283–287, 1927.

[2] Personal communication.

[3] Hintze, C., *Handbuch der Mineralogie* II. p. 254.

In the Santa Rita mountains south of Tucson, Arizona, the writer has found piedmontite in two localities some fifteen kilometers apart. The first found in 1933 was in a nearly white pebble of rhyolitic composition found in an arroyo about six kilometers from the Helvetia copper mines, on the Helvetia-Vail road. The microscope revealed in a thin section only three or four spots of piedmontite in a rock of felsitic texture. The pebble must have been transported several kilometers and no information is available as to its source. The other occurrence in the Santa Rita Mountains was in a dull reddish gray rhyolitic rock in a small canon leading from the east into Madera Canon. The rock was a large angular boulder associated with many others of the same composition and had been transported but a short distance. In fact it probably had broken from the sides of the canon which were also of the same general composition. The piedmontite occurred in a rounded inclusion, of about 10 mm. in diameter, but of unknown original composition. This was almost completely replaced by piedmontite and sericite, the latter being in rather long plates. The area immediately surrounding the inclusion also contained a few inconspicuous grains of piedmontite as well as what appeared to be specular iron. Neither one of these specimens was known to contain piedmontite at the time of discovery but later investigation in the Madera Canon locality proved that specimens of piedmontite-bearing rhyolite can quite easily be secured, the material being limited to loose blocks. The piedmontite then appears as spots or segregations in the rhyolite rock, the matrix itself being rather coarser grained than the similar occurrence in the felsitic rocks of the Tucson Mountains.

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[4] Rosenbusch, H., *Elemente der Gesteinslehre*, 4te Auflage (Osann), pp. 697, 744, 1923.

[5] Williams, G. H., Piedmontite and Scheelite from the Ancient Rhyolites of South Mountain, Pa.: *Amer. Jour. Sci.* (3), vol. **46**, pp. 50–57, 1893.

[6] Harworth, E., *Am. Geologist*, vol. **1**, p. 365, 1888.

[7] Rogers, A. F., Notes on Rare Minerals from California: *School of Mines Quart.*, vol. **33**, p. 373, 1912.

[8] Mayo, E. B., Two new Occurrences of Piedmontite in California: *Am. Mineral.*, vol. **17**, p. 238, 1932.

[9] Short, A. M., A Chemical and Optical Study of Piedmontite from Shadow Lake, Madera Co., California: *Am. Mineral.*, vol. **18**, p. 493, 1933.

In the Tucson Mountains piedmontite occurs in a series of steep quartzite or sandstone peaks and ridges about 1½ kilometers in length extending in a NW to SE direction in what is known as the Tucson Mountain Recreational Area. Its most northly end begins almost exactly at the picnic grounds known as Juan Santa Cruz near the Recreational Headquarters. While the flanks of the peaks and ridges are mainly quartzite or sandstone, some of the summits and crests of the ridges are occupied by a dense pinkish rock of rhyolitic composition, perhaps best designated as felsite. The intrusion of this volcanic mass is probably the direct cause of the sandstone hills. Where the summits are occupied by sandstone perhaps represent places where the intrusion was unable to break through, or else where it has not been exposed by erosion. The felsite is usually very close grained, sometimes breaking like chert into fragments with conchoidal fracture and sharp flint-like edges (Fig. 7). A few phenocrysts of feldspar a millimeter or so in cross section are present but usually very inconspicuous. Its micro-texture, then, seems to prove it to be a cryptocrystalline mixture of quartz and feldspar with frequent clusters of purer quartz much of which is probably a later segregation in the intrusion. Minute veinlets of irregular shape consisting of the same type of quartz are also frequent. The felsite contains inclusions of other volcanics apparently more basic than itself, as well as sandstone probably picked up as it broke through the sandstone series. As the sandstone in places is extremely rich in such volcanic inclusions, the felsites may have obtained theirs from that source.

Almost everywhere in the area under discussion the felsite or rhyolitic rock contains spots of piedmontite clusters of a fine red color when viewed in a strong light. When occurring in larger clusters not mingled with too much clear quartz, the color of the piedmontite is a deep purplish red or brown.

The quartzite or sandstone with which the piedmontite felsite is associated varies in color from buff to dark brown and gray. It also varies greatly in structure from an even fine grained rock, sometimes showing stratification, to a coarse type breaking with a rough surface. In thin sections it often appears with subangular quartz grains of ununiform sizes with a varying amount of cement. In many places, especially near the felsite areas, it is highly arkosic containing many grains of orthoclase, plagioclase and microcline with a cement much resembling the groundmass of quartz porphyries. Indeed it is sometimes so altered as to resemble in thin sections, as well as hand samples, a porphyrite rhyolite or even an aplitic rock. The angular appearance of the quartz, though sometimes lacking due perhaps to resolution, as well as field relations furnish the best criteria. Moreover many volcanic inclusions are present in places but not at all evenly distributed. They consist mainly of rounded pebbles of a porphyritic rock probably of andesitic composition and often in sufficient quantities to give the rock a highly conglomeratic appearance. Thin sections of this sort of material, especially where it is highly silicified and otherwise altered, exactly resemble igneous rocks. It might even be called by some a volcanic breccia but it is believed really to belong to the sandstone series.

The occurrence of the piedmontite in the felsite-sandstone area of the Tucson Mountains may perhaps best be described under the following six types:—

(1) As spots or rounded and lenticular segregations in felsite.

(2) As minute veinlets in felsite a millimeter or two in width and a few centimeters in length.

(3) As replacements of inclusions in felsite, or simply as borders surrounding them.

(4) In veins or stringers of white quartz in both felsite and sandstone, the latter being the better developed.

(5) Associated with calcite in a manner somewhat similar to (4) but of rare occurrence.

(6) As replacements of some of the constituents of sandstone,

and volcanic inclusions therein contained, in the vicinity of the vein system.

The first mode of occurrence is illustrated in Figs. 1, 2, and 3. These are clusters of piedmontite crystals showing in thin sections the usual beautiful trichroism, mingled with quartz which seems to be characteristically segregated near the center of the spot. In Figs. 1 and 2 the less well crystallized borders contain considerable common epidote. These spots vary in size up to about ten millimeters in cross section. Sometimes the crystals of piedmontite

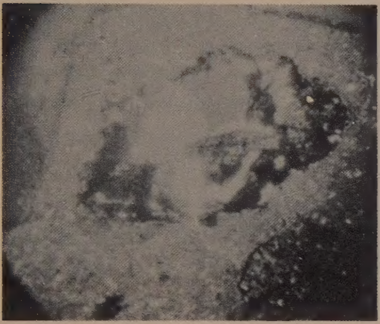

Fig. 1. Thin section of felsite, nicols partially crossed, showing a piedmontite-quartz cluster. ×10.

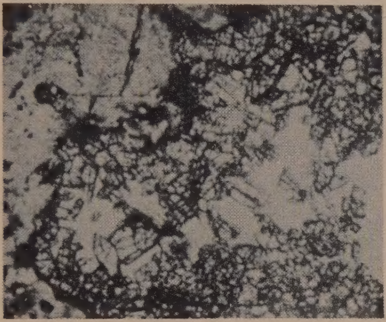

Fig. 2. A larger cluster of piedmontite in felsite. White is quartz, dark with cleavage is piedmontite and mottled with large fracture is felsite. ×10.

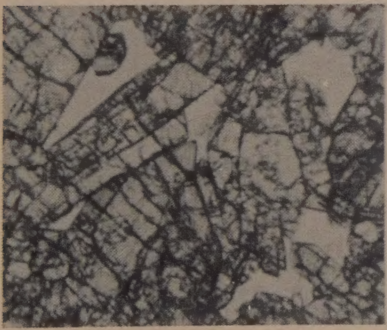

Fig. 3. A photograph of the same as Fig. 2 but with higher power to show texture of piedmontite crystals. Ordinary light. ×36.

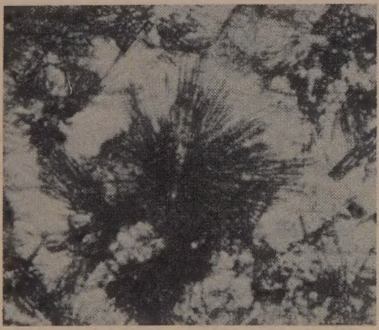

Fig. 4. An illustration to show the frequent stellated habit of piedmontite. White is quartz. Ordinary light. ×34.

when extending towards the center into quartz may be "terminated" by two prominent faces which, as the crystals are usually elongated in the direction of the *b*-axis, are taken to be in the prism zone. Several of these have a normal angle of near 116° as measured by the microscope. Sometimes in place of the spots there are cavities lined with quartz, piedmontite and rarely hyalite, together with such decomposition products as manganese oxide and a kaolin-like mineral. These segregations are thought to represent a concentration of manganiferous material from the body of the felsite itself rather than a migration of manganese from extraneous sources. As discussed later the felsite notwithstanding the fact that it is piedmontite-bearing in practically every place throughout the area, is not abnormally rich in manganese. After the period of vulcanism, some undetermined feature of the hydrothermal solutions segregated the manganese in the form of the epidote molecule from the not unusual traces already present. The felsite as well as the associated sandstone contain numerous black grains, like black sand. Some are magnetic but many are not. They do not react appreciably for manganese.

The second mode of occurrence, that of minute veinlets in felsite, is not materially different from the first except that it emphasizes the fact that the piedmontite has been a later segregation although not necessarily brought in from outside sources. The veinlets may be wandering or they may follow a definite fracture plane. In the latter case the rock may cleave along this plane yielding a surface of pleasing reddish color. This appearance may be enhanced if the surface contains numerous minute quartz crystals. These surfaces show sometimes small red slickensides with striations. On making thin sections of these veinlets as well as the piedmontite spots, the operator is sometimes surprised to find the specimen much leaner in the red mineral than he expected. The minute quartz crystals when backed by piedmontite are frequently taken for the manganese mineral.

The third type of occurrence or the more or less complete replacement of inclusions in felsite is quite common in the Tucson Mountain area. The inclusion may be more or less spotted by the red mineral or crossed by veinlets sometimes forming a net-work. The most characteristic procedure, however, seems to be that of replacements around the border, the piedmontite appearing as rims completely surrounding the inclusion. This type should not be confused with the brecciated inclusions in quartz veins to be described later in which there are still more striking arrangements of the piedmontite as narrow fringes (Figs. 11 and 12). Feldspars in the volcanic inclusions in the felsite have been observed in thin sections partially replaced by piedmontite. Often the whole inclusion has been replaced as described above from material from the Santa Rita Mountains, and there remains only piedmontite and other less rare replacing minerals.

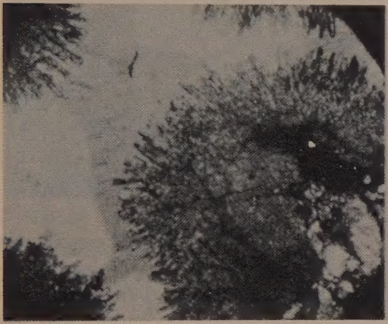

FIG. 5. Fringes of piedmontite on inclusions in quartz veins. Nicols partially crossed. Only a portion of the inclusion can be seen at the lower right. See Figs. 11 and 12 for better display of inclusions and piedmontite fringes. ×25.

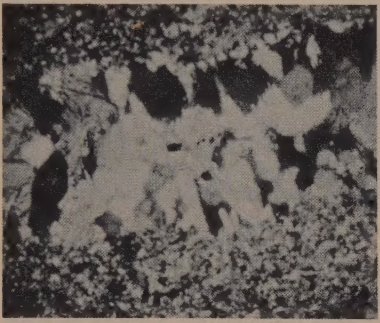

FIG. 6. Veinlet of piedmontite and quartz in sandstone. Note stellated clusters of piedmontite extending from borders of vein into quartz. Nicols partially crossed. ×12.

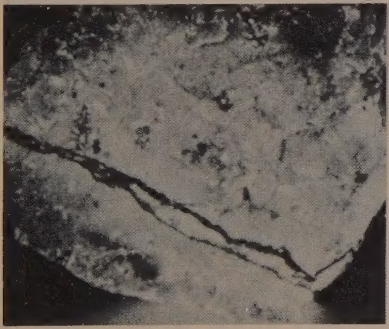

FIG. 7. A photograph of a megascopic specimen of piedmontite-quartz veinlet in dense felsite. Dark borders are piedmontite. ×0.8.

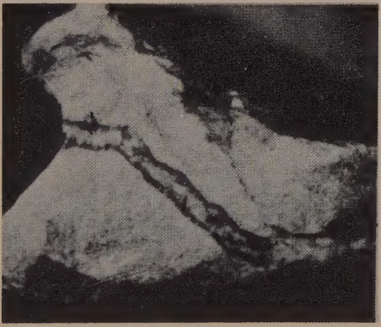

FIG. 8. A similar veinlet of piedmontite and quartz in sandstone still showing the dark borders of piedmontite. ×1.0.

The fourth type, perhaps the most interesting of all, is that of a series of veins starting in the felsite and extending into the sandstone. This occurrence has been well developed in only one place in the Tucson Mountain area, and is illustrated in thin section in

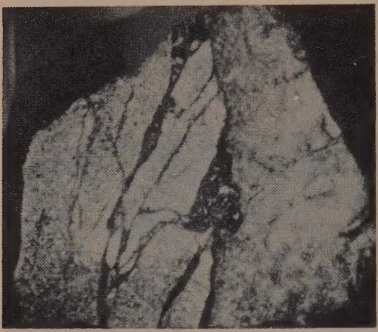

Fig. 9. A photograph of a megascopic specimen showing quartz vein material at right, with country rock (sandstone) crossed by numerous veinlets of piedmontite to the left. Note a few inclusions in the vein material with borders of piedmontite. Such inclusions are better illustrated in Figs. 11 and 12. ×0.7.

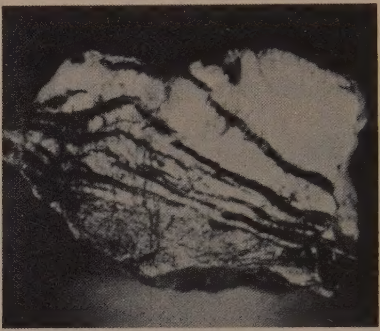

Fig. 10. Another view of quartz vein material, showing gashes of later piedmontite. This is purer than other types found in the Tucson Mountains and was chosen for the chemical analysis. Note portion of wall rock at bottom. ×0.7.

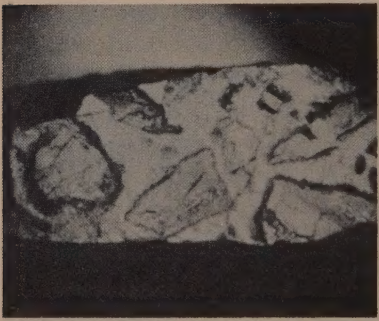

Fig. 11. A typical view (megascopic specimen) of the wider portion of one of the quartz veins, showing inclusions of the wall rock all of which are completely bounded by a narrow fringe of piedmontite. For details in the texture of the fringes see Fig. 5. ×0.4.

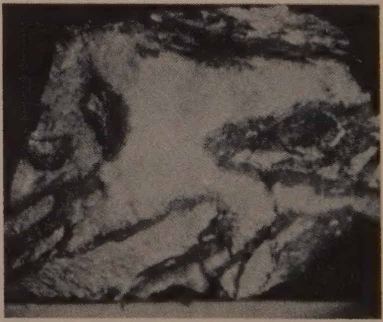

Fig. 12. Same as Fig. 11 but showing inclusions in white vein quartz crossed by piedmontite and, in fact, more completely replaced. ×0.7.

All illustrations accompanying this paper are from the Tuscon Mountains, Arizona.

Fig. 6 and in megascopic specimens in Figs. 7–12. The vein is really a series of narrow stringers of white quartz and piedmontite originating in felsite. Any one stringer may attain a width of 12 cm. while the series may be a meter in width, and is traceable for 25 meters or more. They are exposed on the face of two small cliffs one above the other the lower consisting of sandstone and the upper of felsite. Large blocks broken from the face of the cliffs have yielded the best specimens. The larger white stringers of quartz appear very prominently spotted due to the presence of many subangular inclusions from the sandstone series. These spots may vary from minute fragments to those several centimeters in cross section. The piedmontite appears as striking fringes on practically all of these inclusions. These are shown in photographs of megascopic specimens in Figs. 11 and 12. If the inclusion is small or if the break crosses a narrow portion of it, the spot may be made up entirely of piedmontite. On weathered surfaces the inclusions have all taken on a rather uniform purplish black color presenting a strong contrast to the white quartz. On freshly broken surfaces the fringes of piedmontite correspond to Vandyke Red 1′ Red k,[10] and may vary from one to several millimeters in thickness. Yet when examined by a hand lens in a strong light or in thin sections the color is that of a rich ruby red. A photomicrograph of some of these fringes together with a small portion of the inclusion is shown in Fig. 5. The quartz stringers may also have later gashes of purer piedmontite of considerably darker color or near Dark Mineral Red 1′′ Red m, the difference in shade being due to the amount of quartz present. These piedmontite gashes may reach a width of two centimeters and being freer from impurities than the fringes on inclusions or the spots in felsite were chosen for the selection of material for the analysis given in the table.

Silica, lime and alumina correspond rather closely to material analyzed by Takayama[11] while manganese in considerably higher, but the sum of iron and manganese sesquioxides is almost identical. Manganese is only slightly lower than in piedmontite from California recently analyzed by Kameda[9] but higher than the partial analysis (8.91%) by Dr. Buehrer in piedmontite from Sulphur Spring Valley, Arizona.[1] Manganese is lower than in six of the seven analyses in Hintze's Handbuch. The RO-R_2O_3-SiO_2 ratio corresponds to the generally accepted figures (4-3-6) except that lime is rather high.

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[10] Ridgway, R., *Color Standards and Nomenclature*, 1912.

[11] Hintze, C., *Handbuch der Mineralogie*, II. p. 256, (Anal. VII).

ANALYSIS OF PIEDMONTITE, TUSCON MOUNTAINS, PIMA CO., ARIZONA

Dr. Paul H. M.-P. Brinton, Analyst

Silica, (SiO_2)	37.43%
Alumina, (Al_2O_2)	21.27
Iron Oxide, (Fe_2O_3)	3.80
Calcium Oxide, (CaO)	24.75
Magnesium Oxide, (MgO)	0.00
Water, Below 110°	0.04
Water, Above 110°	0.92
Manganese Sesquioxide, (Mn_2O_3)	11.80
Titanium Oxide, (TiO_2)	0.10
Total	100.11

Material from the same sample as analyzed by Dr. Brinton was selected for determination of refractive indices and trichroism with the results given below:—

REFRACTIVE INDICES OF PIEDMONTITE FROM TUCSON MOUNTAINS, ARIZONA

By Dr. Esper S. Larsen

$\alpha=1.732$	Buff, 15′b. Ridgway's Color Standards.
$\beta=1.750$	Deep Vinaceous Lavender, 65‴d.
$\gamma=1.778$	Spinel Pink, 71′b.

These values are lower, β and γ considerably lower, than suggested by the curves showing the relation of manganese content and refractive indices published by A. M. Short.[12] These curves, however, contain but two determined points.

In thin sections this dark piedmontite occurring as gashes in quartz is sometimes seen to grade into a very fine grained type consisting of fine needles or fibers which further grade into material so fine as to be nearly isotropic and without pleochroism. This may pass into wavy varieties resembling agate structure but observable only in thin sections.

A few of the thicker gashes of piedmontite contain a dullish metallic mineral which from its hardness and other mineralogical tests was taken to be braunite. This is reasonably confirmed by a partial analysis by Dr. Brinton. The small amount of material available was too impure to make a complete analysis worth while. Dr. Brinton's results are as follows:—

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[12] Short, A. M., A Chemical and Optical Study of Piedmontite from Shadow Lake, Madera County, California; *Am. Mineral.*, vol. **18**, p. 498, Fig. 4, 1933.

Total Silica (SiO_2)	26.82%	
Manganese Sesquioxide (Mn_2O_3)	53.18	
Insoluble residue		34.8%

The insoluble residue was largely piedmontite with quartz. A complete analysis of piedmontite from the same material made by Dr. Brinton and described above contained 37.43% silica. Therefore of the 34.8% insoluble residue, 37.43% is, disregarding other impurities, silica combined in piedmontite. This leaves 13.8 parts for silica associated with manganese sesquioxide. On recalculating to 100% we have for the braunite 79% manganese sesquioxide and 21% silica. Polished surfaces show that the braunite is roughly oriented in bands parallel to the elongation of the piedmontite gashes. It was associated with sharp acicular crystals, probably an oxide of manganese.

The above description of the vein material refers to that in the sandstone series. That portion of the vein system in felsite is somewhat different, the stringers being narrower and the quartz veins lacking the prominent brecciated condition and the resulting inclusions. Many of the stringers show striking borders of piedmontite, the centers being mainly occupied by white quartz (Fig. 7). Some of the very narrow veinlets, say one or two millimeters in width, may appear in hand specimens as all piedmontite, yet on examination in thin section are found really to have quartz in excess. The admixture of quartz seems to be the chief cause for the variability in the shade of red. The piedmontite appears as tufts of divergent crystals extending from the walls of the veinlet. Fig. 6 in thin section and Fig. 8 a photograph of a megascopic specimen show this same type in sandstone.

The fifth type of occurrence is piedmontite associated with calcite. This occurs a short distance from the quartz veins and resembles them except calcite is found instead of quartz. The formation is rather insignificant and only a few specimens are available. On examination with a hand lens the characteristic tufts of piedmontite seem to be entering the calcite in the same manner as the corresponding tufts in the quartz veinlets. On investigation in thin sections, however, the tufts are found to be distorted the separate needles being curved and many times broken and filled in by calcite. From this it is taken that calcite is later than the piedmontite and has simply replaced the quartz in the veinlet which is held to be earlier than piedmontite.

The sixth mode of occurrence, that of scattered grains of piedmontite in sandstone, probably occurs only in the vicinity of the piedmontite-quartz veins. The stray grains are about the size of the constituents, and have been observed in thin sections up to five meters from the vein. Sometimes it is paler here and associated with and grading into a near colorless epidote.

Many of the specimens described above show on weathered surfaces a thin film or black (purplish) varnish either on the piedmontite itself or the associated rock. This reacts strongly for manganese and is without doubt an oxide of manganese. Much of this does not differ particularly from the well known "desert varnish" described by Blake.[13]

Piedmontite has been described as sometimes having red centers or ends grading off into colorless or greenish borders, a phenomenon to be expected when its probable isomorphism with common epidote is considered. Lausen,[1] has described this feature in material from Pat Hills, Arizona. This is not a particularly noticeable feature in the Tucson Mountain area. A few sections do show a deep red center with an abrupt change to a colorless border, while other grains usually associated with common epidote have but the faintest pink tint.

ORIGIN OF PIEDMONTITE FROM THE TUCSON MOUNTAINS. In the discussion of the origin of common epidote little attention need be paid to the source of the iron or other elements as they are present in notable quantities in all rocks. Naturally the source of the iron is held to be from the breakdown of the ferromagnesian minerals such as biotite, hornblende, etc. The situation seems to be somewhat different in the case of piedmontite as there are no typical manganese rock-forming minerals. Thus it might be held that by some process, not easily traceable, these piedmontite rocks became unusually rich in manganese. A determination of manganese was therefore made in the rocks of the piedmontite area including piedmontite bearing felsite and sandstones as well as two normal rhyolites of the Tucson Mountains, but outside the piedmontite district. The results are given in the table below.

Perhaps the most interesting feature brought out in the above table is the manganese content of the piedmontite felsite as shown by the first three determinations. The percentages, here calculated as Mn_2O_3, are not above those frequently reported everywhere in normal rhyolitic rocks, yet in all three piedmontite could easily be detected by a hand lens. This shows that a rock, in order to develop piedmontite does not necessarily have to contain more manganese than is found in normal rocks almost anywhere. The development of the piedmontite then must be due to some undetermined peculiarity of the geophysical conditions. It may be that these conditions are a little different than those required to form common epidote. It is therefore believed that the manganese which has given rise to piedmontite has been present in the body of the felsite and search for an extraneous source is not necessary.

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[13] Blake, W. P., Superficial Blackening and Discoloration of Rocks Especially in Desert Regions: *Am. I. M. Eng.*, vol. **35**, pp. 371–375, 1905.

Percentages of Manganese in Piedmontite Felsite and Associated Rocks from Tucson Mountains, Arizona

Dr. Paul H. M.-P. Brinton, Analyst

No. 1.	Piedmontite Felsite, south end	0.14% Mn_2O_3
No. 2.	Piedmontite Felsite, near center	0.07
No. 3.	Piedmontite Felsite, north end	0.07
No. 4.	Conglomerate sandstone, south end about 8 meters from a felsite intrusion	0.21
No. 5.	Sandstone, even grained and showing piedmontite under microscope, 5 meters from piedmontite vein	0.05
No. 6.	Sandstone, about 300 meters from piedmontite felsite	0.25
No. 7.	Sandstone, north end	0.15
No. 8.	Pink rhyolite cliffs, about 3½ kilometers from the piedmontite felsite area	0.05
No. 9.	Normal rhyolite, Tucson Mountains, Amole Peak	0.06

Manganese was determined in four samples from the sandstone series (Nos. 4–7) distributed as indicated in the table. Rather varying amounts were found as might be expected from the variable character of the sandstone. Those remote from the piedmontite veins seemed to carry fully as much manganese as those near by, perhaps the richest being those with large quantities of volcanic inclusions. Although the sandstones are higher in manganese than the piedmontite felsites, they cannot be looked upon as the source of manganese for the genesis of the piedmontite, as it seems irrational to formulate the necessary movement of the hydrothermal solutions. Where piedmontite is found in sandstone it has been assumed that it has resulted from migrations from the felsite. Some geophysical condition associated with post-volcanic activity was the great factor in the development of piedmontite wherever found.

Manganese was determined in two rhyolites considerably removed from the piedmontite area (Nos. 8 and 9) and these were found to contain about the same amount of Mn_2O_3 as two out of three samples of piedmontite felsite. No. 8 was strikingly pink in color and it was thought for that reason it might contain piedmontite, but several thin sections failed to reveal any. No. 9 was a light gray rhyolite and represented, as compared to the felsite area a very large mass. This determination was made in order to compare the manganese content in the piedmontite felsite with normal rhyolite of the Tucson range of mountains.

RESUMÉ

(1) One new locality for piedmontite in Arizona has been recorded, that of the Santa Rita Mountains south of Tucson, where it is found in spots in rhyolitic rock and replacing inclusions. The description of this has been brief as only few data are available at the present time.

(2) A detailed account of the occurrence of piedmontite in felsitic rock of the Tucson Mountains together with one new complete analysis and optical determinations, make up the most important part of the paper. Six types of occurrence have been described.

(3) The richest specimens of piedmontite are found where the hydrothermal solutions have converged in the form of a series of white quartz stringers into sandstone. These veins contain numerous angular inclusions of the country rock which are, almost without exception, fringed with striking layers of purplish red piedmontite. The smaller veinlets may consist of white quartz centers and piedmontite borders.

(4) The felsite itself has been considered the source of the manganese from which piedmontite has been segregated into spots and the other types of occurrence. The felsite as a whole does not seem to contain, as shown by a chemical determination, more manganese than is frequently found in normal rhyolitic rocks.

(5) The complete analysis of piedmontite, a partial analysis of impure braunite, and numerous determinations of manganese for a discussion of the geological distribution of that element in the Tucson Mountains, were made by Dr. Paul H. M.-P. Brinton, of Pasadena, California, and the determination of the indices of refraction and pleochroism were made by Dr. Esper S. Larsen of Harvard University. For this valuable assistance the writer is pleased to express his sincere appreciation.

SPHERULITIC CONCRETIONS OF DAHLLITE FROM ISHAWOOA, WYOMING

Duncan McConnell, *Stanford University, California*

During the summer of 1934 some unusual concretions were shown to the writer but they were not examined in place because their mineralogical nature was not fully appreciated until they were later examined in the laboratory, and because their location in the field was inaccessible. They occur in abundance in the basal part of the Thermopolis formation (Upper Cretaceous), and are useful as stratigraphic markers in the region near Cody, Wyoming.[1] The ones examined by the writer were collected about ten miles southwest of Ishawooa, Park County, Wyoming.

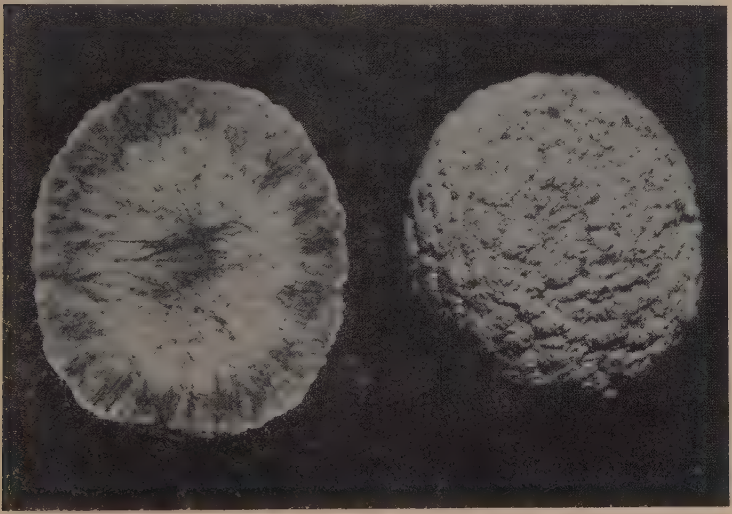

Fig. 1. Polished cross section showing dark center and very narrow border of incipient weathering. The peculiar external surface is shown. ×4/3.

Megascopic Observations

Some of the concretions are compound, i.e., composed of several interpenetrating spheres. Many, however, are approximately spherical in shape and average about 3 cm. in size. They are phosphatic nodules and are unusual, first, in their spherulitic structure

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[1] Private communication from Mr. E. H. Stevens, of the University of Chicago.

(Figs. 1 and 2) and secondly, in consisting essentially of dahllite which, although probably a frequent constituent of phosphate nodules, generally does not predominate. The exterior has a peculiar texture, which usually shows a slight amount of weathering (Fig. 1). The color of the weathered surface is light gray. Immediately within the weathered zone the material is brown, grading inwardly to light yellowish-gray. The center of the concretions is different in texture and composition; it is very dark in color.

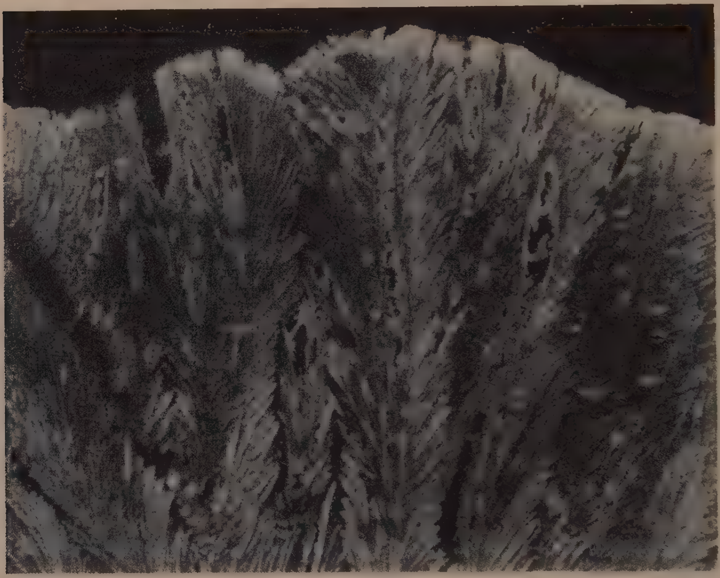

Fig. 2. The plumose fibers of the spherulitic structure containing numerous small cavities and filled cavities. (Oblique reflected illumination.) ×8.8.

The dahllite effervesces in cold dilute acid and gives water in the closed tube, besides producing microchemical tests for phosphate and calcium ions.

Microscopic Observations

The spherulitic concretions are composed of two types of fibers (Fig. 3). The more apparent fibers are the broad type and these are aggregates of well defined microcrystals whose elongation is transverse to the length of the fibers. The more predominant, the narrow type, are composed of microcrystals whose length is parallel to the length of the fibers. The microcrystals of these fibers are poorly defined and the fibers are interrupted throughout their length. In both types of fibers the optical orientation of the microcrystals is length-fast.

When the sensitive red plate is inserted between crossed nicols in a position such as to cause the retardation produced by the broad type of fibers to increase, the retardation produced by the narrow type decreases, while still other portions remain at extinction. The portions at extinction are intimately associated with the fibers of the narrow type and seem to indicate a poor orientation of the microcrystals composing them.

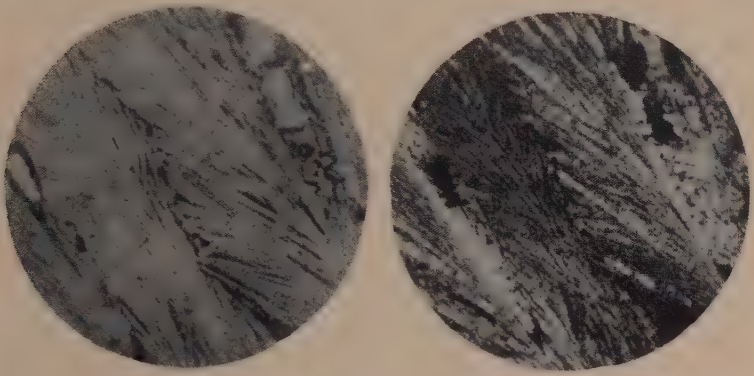

FIG. 3. Fibers, cavities, and small patches of the dark material which forms the center. ×22. Same with crossed nicols showing narrow and broad types of fibers, spherulites, and partial spherulites. ×18.

Very minute secondary spherulites and partial spherulites occur at the margins of numerous small cavities or within a cryptocrystalline material which has filled some of the cavities (Fig. 3). Associated with the filled cavities are aggregates of small euhedral crystals of dahllite (average 30μ–40μ in length—maximum 70μ), which are prismatic in habit and show well defined hexagonal basal sections (10μ across). The cryptocrystalline ground mass filling the cavities varies from colorless to yellowish-brown. Its composition is not known but it is probably a member of the kaolin group.

The small euhedral crystals have γ parallel to c and refractive indices:

$$n_\alpha = (\epsilon) = 1.627\ (\pm 0.003)$$
$$n_\gamma = (\omega) = 1.635$$
$$\Delta = 0.008\ (\pm 0.002)$$

The only observed forms were the prism and pinacoid.

The dark material in the center of the concretions (Fig. 1) is translucent in thin sections and contains iron and a small amount of manganese, as indicated by microchemical and blowpipe tests, but it cannot be referred to any mineral species and probably represents a mixture of oxides. Within this material calcite occurs, filling interstices, and in the dahllite, calcite occurs at the margins of cavities to a lesser extent. The dark material seems to be later than the dahllite and the calcite is probably still later.

Examination of a polished surface disclosed that minute crystals of pyrite occur in the dark material of the center and also in the dahllite. Most of these specks of pyrite show alteration and many of them are cross-cut by stringers of oxides.

No organic remains (sponge spicules, bones, shell fragments, etc.) were found in the thin sections. This suggests that the fibrous dahllite did not directly replace organic matter. The concretions may represent replacement of some other material but, if this be the case, the replacement was quite complete, because no relicts of the replaced material were found. It is possible that the dahllite represents a crystallization product of amorphous calcium carbonate-phosphate, and it is further possible that collophane is intermixed with the microcrystals of the narrow fibers, although none could be detected with certainty.

NOTES ON DAHLLITE

The apatite group may be considered as being composed of:

fluor-apatite	$Ca_{10}F_2(PO_4)_6$
chlor-apatite	$Ca_{10}Cl_2(PO_4)_6$
hydroxy-apatite	$Ca_{10}(OH)_2(PO_4)_6$
oxy-apatite or voelckerite	$Ca_{10}O(PO_4)_6$
carbonate-apatite or dahllite	$Ca_{10}CO_3(PO_4)_6$

The above formula is that accepted for dahllite by Rogers,[2] Eitel,[3] and Brauns.[4] With the addition of a combined molecule of water it is the one accepted by Washington,[5] and by Schaller.[6] Other authors accept either the original formula, $2Ca_3(PO_4)_2 \cdot CaCO_3 \cdot \frac{1}{2} H_2O$, or one of several slight modifications involving the amount of water. In general, those who contend that the calcium phosphate to calcium carbonate ratio is two to one consider that podolite is a distinct species, while those who consider the three to one ratio as correct for dahllite regard podolite as identical with dahllite. The probability of error in determining the small amount of fluorine, which is frequently present, together with lack of concordance in the determinations of water permit sufficient margin for the differences of opinion.

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[2] Rogers, A. F., Dahllite (podolite) from Tonopah, Nevada, voelckerite a new calcium phosphate; remarks on the chemical composition of apatite and phosphate rock: *Am. Jour. Sci.*, vol. **33**, pp. 475–482, 1912.

[3] Eitel, W., Über Carbonatphosphate der Apatitgruppe: *Schr. Königsb. gelehrt. Ges., Naturw. Kl.*, **1924**, pp. 159–177.

Dahllite usually occurs in phosphorite, as fibrous incrustations frequently showing colloform structure, where it is commonly associated with collophane. This is the mode of occurrence in Quercy.[7] It has been reported as a secondary mineral filling small fractures in feldspar-poor alkali basalt in Odenwald[8] and in naujaite in Greenland.[9] At the locality from which dahllite was originally described it was associated with apatite.[10] It occurs at Tonopah, Nevada, in the oxidized zone associated with iodyrite, hyalite, quartz and manganese dioxide.[11] At Cerro Mercado, near Durango, Mexico, dahllite is associated with martite and goethite.[12] It is also occasionally found in fossil bone.[13]

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[4] Brauns, R., Über den Apatit aus dem Laacher Seegebiet, Sulfatapatit und Carbonatapatit: *N. Jahrb. Miner., B. B.*, vol. **41**, pp. 60–92, 1917.

[5] Washington, H. S., Dahllite from St. Paul's Rocks (Atlantic): *Am. Mineral.*, vol. **14**, pp. 369–372, 1929.

[6] Schaller, W. T., The probable identity of podolite with dahllite: *U. S. Geol. Surv., Bull.* **509**, pp. 96–97, 1912.

[7] Lacroix, A., Sur la constitution minéralogique des phosphorites français: *Compt. Rend.*, vol. **150**, pp. 1213–1217, 1910. Lacroix, A., *Minéralogie de la France,* vol. **4**, pp. 555–558, 1910.

[8] Nieland, H., Carbonatphosphate (Dahllit, Quercyit usw.) vom Katzenbuckel im Odenwald: *N. Jahrb. Miner., B. B.*, vol. **59, A**, pp. 205–222, 1929.

[9] Böggild, O. B., Dahllit från Kangerdluarsuk: *Meddel. om Grönld.*, vol. **51**, pp. 434–443, 1915; also *Z. Krist.*, vol. **55**, pp. 417–425, 1920.

[10] Brögger, W. C., and Bäckström, H., Über den Dahllit, ein neues Mineral von Ödegården, Bamle, Norwegen: *Öfv. Akad. Stockh.*, vol. **45**, p. 493, 1888.

[11] Rogers, A. F., *Op. cit.*

[12] Foshag, W. F., Mineralogy and geology of Cerro Mercado, Durango, Mexico: *U. S. Nat. Mus., Proc.* **74**, art. 23, p. 12, 1928.

[13] Rogers, A. F., Mineralogy and petrography of fossil bone: *Geol. Soc. Amer., Bull.* **35**, p. 549, 1924.

The refractive indices of dahllite have been reported to vary from $n_\alpha = 1.591$–1.631 and $n_\gamma = 1.598$–1.635 with $n_\gamma - n_\alpha = .004$–$.019$. Although the mineral is usually uniaxial negative it has been reported as showing optical anomalies. The habit may be either prismatic or tabular; showing, most frequently, the simple prism and pinacoid but also the unit bipyramid.

The mineral collection of Stanford University contains phosphate concretions from Podolia, U.S.S.R., and comparisons of hand specimens and thin sections have shown them to be somewhat similar to those from Wyoming. The concretions from Russia, however, are larger, darker in color, have a worn exterior surface, show a lesser development of the plumose structure and contain glauconite, quartz, and chalcedony intermixed with the fibers and anhedral crystals of dahllite. It is probable that these concretions may have come from the locality from which podolite was described.[14]

Although the concretions from Wyoming are somewhat similar to those found in Podolia they exemplify a rare type of concretionary structure, and this occurrence of dahllite in the United States seems noteworthy.

ACKNOWLEDGEMENTS

The writer wishes to acknowledge the kindness of Mrs. F. C. Sayles, Jr., who called to his attention these concretions and who supplied the specimens, and of Professor A. F. Rogers, who supplied specimens of dahllite for comparison and gave constructive criticism of the work.

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[14] Tschirwinski, W., Über Podolit, ein neues Mineral: *Centrb. Mineral.*, **1907**, pp. 279–283.

THE STRUCTURAL RELATIONSHIP OF GLAUCONITE AND MICA

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INTRODUCTION

The composition, properties and origin of glauconite have been investigated extensively in the past, but no attempt has been made to determine its structure. The writer in following up his work on other layer structures *x*-rayed eight glauconites under a number of different conditions and arrived at the conclusion that glauconite is a mica in structure. He is indebted to Dr. Clarence S. Ross and Dr. W. F. Foshag for gifts of analyzed and unanalyzed samples of glauconite. Mr. Gilman Berg was of great assistance in purifying some of the samples by the dielectric method. Generous grants from the Graduate School of the University of Minnesota have made this and previous structure studies possible.

Glauconites used in the investigation were:

1. St. Francis County, Missouri. From the St. Joseph Lead Company mine, near Bonneterre. In the Bonneterre dolomite (upper Cambrian). Analysis, Table 3, as given by C. S. Ross (1, p. 10). Optical properties by Ross: negative, $\alpha=1.597$, $\beta=1.618$, $\gamma=1.619$; $\gamma-\alpha=0.022$, $2V=20°$, $2E=33°$. The acute bisectrix X is nearly but not quite normal to the basal cleavage, which is good. The absorption is $Z=Y<X$, pleochroism Z and Y lemon yellow, X dark bluish-green.

2. Huntington, Oregon. This is an unusual glauconite of which Ross (1, p. 4) says:

> It forms large compact masses of an earthy texture. In thin section it resembles massive serpentine, with large, poorly defined, smearlike areas of birefracting material. No sharply defined crystals were observed and the cleavage is not well developed.

$\alpha=1.59$, $\gamma=1.62\pm.005$, $2V=20°\text{–}40°$, negative. Analysis, Table 3, quoted from paper by Ross.

3. Southeastern Minnesota. Franconia formation (upper Cambrian).

4. Black Hills, South Dakota. Deadwood formation (Cambrian).

5. Mobile County, Alabama (Eocene).

6. New Brunswick area, New Jersey (Eocene). United States National Museum No. 97761.

Two others of unknown origin.

X-RAY DATA

From the beginning of this study it was considered highly probable, based on previous optical data, that glauconite is very similar to the layer structures of the micas, kaolinites, vermiculites, or chlorites. This view had been held by a number of mineralogists for years. On account of the nature of the material the powder method of *x*-ray analysis provided the only means of attack. Throughout the work Fe K radiation from a modified Ksanda ion tube was used. Exposures varied from 20 to 50 hours at an average of 7 ma. at 30 kv. The radii of the cameras were 57.3 mm. All powders were mounted with collodion on silk thread, and the rods thus formed were about 0.8 mm. in diameter. The glauconites were so clean when separated dielectrically that no foreign lines of quartz or other minerals could be detected in the powder spectrum photographs.

As may be seen in Table 1, the interplanar distances *d* of the various specimens are surprisingly uniform, and the relative intensities of the lines (estimated by eye) do not vary appreciably. However, the lines lack the sharpness found in micas of coarser crystallization. The broadening of the lines may be partly due to the extremely small size of the crystal grains of glauconite, which for a large portion of the material must be below the optimum particles sizes for the powder method. They were so fine that no grinding of the samples was necessary beyond simple crushing. Broadening of lines could also be caused by slight distortions in the crystal lattice, modifying it very slightly from monoclinic to triclinic symmetry.

Table 1 gives the dimensions of the unit cell of each specimen and the densities of two of them as determined with the pycnometer.[1] The density of No. 1 is the average of two determinations on about 0.3 gms. No. 6 was determined on about 2.5 gms. of material, which contained a few grains of quartz and limonite. The material from Oregon, No. 2, was not suitable for density determination on account of its exceedingly great porosity and fineness. The other glauconites were not available in large enough pure samples.

If a powder diagram of glauconite were compared with one of muscovite, no similarity probably would be noticed except for the basal reflections 002 and 006, which almost coincide. Even if an iron-rich mica (biotite) pattern is used for comparison the similarity is not conspicuous until allowance is made for the probable difference in the dimensions of their unit cells, and remembering

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[1] To fill all pore space the contents of the pycnometer were boiled under reduced pressure.

TABLE 1. POWDER DIAGRAMS OF GLAUCONITES. FE RADIATION.

No.	1		2		3		4		5		6		Indices	Biotite*	
	d	*I*	*d*	*I*	*d*	*I*	*d*	*I*	*d*	*I*	*d*	*I*		*d*	*I*
1	9.94	1	9.99	2	9.95	1	9.99	1	9.99	1	9.96	2	002	9.95	4
2	4.91	0.5	4.96	0.5	4.99	0.5	4.97	0.5	4.95	0.5	4.93	0.5	$110\beta, 11\bar{1}\beta$		
3	4.49	2	4.43	3*b*	4.53	3	4.52	2	4.48	2	4.53	2	$110, 11\bar{1}$	4.60	0.5
4	3.67	2*b*	3.63	1*b*	3.64	2*b*	3.64*b*	2	3.63	1*b*	3.65	2*b*	$023?, 113?, 006\beta, 024$	3.68	1
5	3.313	3	3.313	2	3.323	3	3.323	2	3.313	2	3.313	3	006, 024	3.333	5
6	3.092	1*b*	3.082	1*b*	3.082	2*b*	3.074	1*b*	3.044	0.5*b*	3.077	0.5*b*	025, 115, 114	3.126	0.5
7	2.859	1	2.836	1	2.849	1	2.852	1	2.845	1	2.843	1	$20\bar{2}\beta, 131\beta, 115, 11\bar{6}?$	2.915	0.5
8	2.681	0.5	2.669	0.5	2.683	0.5	2.669	0.5						2.701	*Ind.*
9	2.582	4	2.573	4	2.582	5	2.566	3	2.573	3	2.574	4	$20\bar{2}, 131, 116$	2.630	2
10	2.402	2*b*	2.392	3*b*	2.399	3*b*	2.392	2*b*	2.384	2*b*	2.394	2*b*	$20\bar{4}, 133$	2.434†	1
11	2.256	0.5	2.248	0.5	2.269	1	2.256	0.5			2.267	0.5	$040, 22\bar{1}, 041$	2.260	0.5
12	2.191				2.214	0.5	2.196	0.5			2.200	0.5		2.196	0.5
13	2.141	1*b*	2.141	1*b*	2.127	1*b*	2.119	1*b*	2.119	1*b*	2.141	1*b*	$20\bar{6}, 135, 043$	2.170	2
14	1.999	1*b*	1.964	1*vb*	1.980	1*b*	1.986	1*b*	1.968	1*b*	1.985	1*b*	$206, 13\bar{7}, 0010$	1.999	2–3
15	1.823	0.5			1.823	0.5	1.827	0.5					$20\bar{1}0\beta, 139\beta, 0012\beta$	1.846†	0.5
16	1.718	0.5	1.703	0.5	1.716	0.5	1.710	0.5	1.711	0.5	1.709	0.5		1.701	0.5
17	1.656	2*b*	1.645	2*b*	1.650	3*b*	1.649	2*b*	1.649	2*b*	1.643	2*vb*	$20\overline{10}, 139, 0012$	1.672	3
18					1.596	0.5*b*					1.590	0.5			
19	1.516	4	1.512	4	1.514	6	1.510	4	1.508	3	1.511	3	$060, 33\bar{1}, 13\overline{11}, 2010$	1.539	4
20	1.501	1	1.495	0.5	1.499	1	1.496	0.5			1.491	0.5	$33\bar{3}, 331$	1.523†	0.5
21					1.445	0.5					1.422	0.5	0014, 335, 333	1.433†	1
22	1.379	0.5			1.375	0.5					1.375	0.5	$33\bar{7}, 335$	1.399	0.5
23	1.337	0.5*b*	1.341	1*vb*	Broad	*Ind.*					1.343	0.5*b*	$2012, 13\overline{13}$	1.363	3
24	1.306	1	1.304	3	1.309	3	1.304	2	1.304	1	1.306	2	$40\bar{2}, 260, 400, 26\bar{2}$	1.330	0.5
25					1.284	1					1.279	0.5	402, 264	1.277†	0.5
26	1.254	0.5	1.254	1	1.259	1	1.250	0.5	1.248	0.5	1.252	1	$0016, 264, 40\bar{6}$	1.255	0.5
27	1.204	0.5	1.203	0.5	1.205	1	1.201	0.5					$2014, 13\overline{15}, 339$	1.218	0.5
c_0	19.96		20.06		20.02		20.02		20.00		19.98			20.11	
b_0	9.10		9.07		9.08		9.06		9.05		9.10			9.23	
a_0	5.25		5.24		5.24		5.23		5.23		5.25			5.33	
β	95°		95°		95°		95°		95°		95°			95° 04′	
S.G.	2.855										2.81			3.151	

Ind. = indistinct edge of line.
b = broad line.
vb = very broad line.

* From Mora, Minnesota.
† Weak lines are also present at $d = 2.503$, 1.910, 1.502, 1.474, 1.310.
β = a line produced by K β radiation of iron.

that weak lines of mica would be absent in glauconite on account of their diffused character, even though the two structures were the same. Another factor to be considered is the preferred orientation of the mica lamellae parallel to the thread, which causes a much greater observed intensity for basal planes. This would not be the case with glauconite because each grain of powder constitutes an aggregate of still smaller particles.*

EXPERIMENTAL DATA

There can be little doubt that next to talc the micas possess the most stable layer structures at high temperatures. Experiments on the stability of glauconites, biotites, and other layer structures were made at 750°C. The heating was carried out in air as well as in CO_2 over periods of twenty hours. Only the glauconites, biotites (other micas were not tested), and the almost iron-free sheridanite (12) did not collapse. Chlorites containing considerable iron, vermiculites, montmorillonite, nontronite, and stilpnomelane broke down under these conditions. Two of the diagrams of glauconite after heating are given in Table 2. They indicate a very slight expansion of the lattice in the direction normal to the cleavage and a corresponding contraction in the direction of the *b*-axis.

Other stability experiments were conducted in steel bombs entirely lined with gold. The glauconite was heated in solutions at temperatures as high as 300°C. over periods as long as seven days. Some of the experiments are given below, and the resultant *x*-ray diagrams are listed in Table 2.

1. No. 2 glauconite at 200°C. in water containing a very slight amount of NH_4F. Five days. Table 2, column 3.
2. The same at 300°C. Table 2, column 4.

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* Except for one line (No. 4, Table I) the agreement would be quite excellent. This line is considerably stronger and broader than any corresponding line in biotite. It also has the appearance as if its outer edge is rather variable in position between 3.62Å and 3.67Å. Such anomalous behavior and intensity of a line, however, are not unique for glauconite. They may be found in other extremely fine-grained crystalline substances, as for example, nacrite, vermiculite, stilpnomelane, and montmorillonite. The finer the grain the more likely they will be. The writer has no explanation to offer for them at present. The biotite which is used for comparison in Tables 1 and 5 is from a granite at Mora, Minnesota (8). Its analysis is given in Table 3. This biotite is one of five analyzed biotites which were used for *x*-ray analysis. It is typical for all of them.

Table 2. Powder Diagrams of Glauconites Subjected to Various Treatments

No.	Glauc. No. 1		2		2		2		2		6	
	750°C.		750°C.		H_2O at 200°		H_2O at 300°		Thallium		Thallium	
	d	*I*	*d*	*I*	*d*	*I*	*d*	*I*	*d*	*I*	*d*	*I*
1	10.14	2	10.17	3	10.13	2	9.99	2	9.99	1	9.99	2
2					4.95	0.5	4.99	0.5				
3	4.48	1	4.49	2	4.50	2	4.51	2	4.48	1	4.53	2
4	3.66	0.5			3.62	1	3.64	1	3.59	1	3.65	0.5
5	3.333	4	3.368	3	3.295	3	3.304	2	3.285	2	3.304	3
6	3.126	0.5			3.066	1	3.066	1	3.052	1	3.074	1
7	2.863	0.5	2.852	0.5	2.836	1	2.838	0.5	2.829	0.5	2.852	1
8			2.702	0.5	2.647	0.5	2.669	0.5	2.669	0.5	2.683	0.5
9	2.571	3	2.577	3	2.562	4	2.573	3	2.551	3	2.568	4
10	2.389	2	2.387	2	2.383	3	2.392	2	2.370	2	2.389	3
11	2.247	0.5	2.248	0.5	2.248	1	2.244	1	2.247	1	2.263	1
12					2.187	1*b*	2.191	1	2.187	1	2.207	1
13	2.142	0.5	2.141	0.5	2.121	1*b*	2.130	1*b*	2.127	1*b*	2.135	1
14			1.968	0.5*b*	1.968	0.5*b*	1.983	0.5	1.962	1*b*	1.985	1
15	1.820	0.5	1.831	0.5	1.812	0.5	1.810	0.5	1.805	0.5	1.817	0.5
16					1.708	0.5	1.697	0.5	1.699	1	1.705	0.5
17	1.655	1	1.653	1	1.637	2*vb*	1.641	1*vb*	1.636	2*vb*	1.641	2*vb*
18					1.586	0.5	1.589	0.5	1.580	0.5	1.589	0.5
19	1.506	2	1.509	3	1.504	4	1.508	3	1.503	3	1.510	4
20			1.492	0.5	1.489	0.5			1.485	1	1.489	1
21			1.439	0.5	1.435	0.5	1.448	0.5			1.428	0.5
22	1.366	0.5										
23			1.355	0.5	Broad	*Ind.*					1.345	0.5*b*
24	1.299	1	1.303	1	1.299	2	1.299	1	1.296	2	1.301	2
25	1.283	0.5					1.270	0.5	1.270	0.5		
26	1.247	0.5			1.249	0.5	1.249	0.5	1.250	0.5	1.252	1
27											1.200	0.5
										Sp. Gr.	3.02	

Ind. = indistinct.
b = broad.
vb = very broad.

3. No. 2 glauconite at 300°C. in $NaHCO_3$ (concentrated solution at room temperature). Nine days. Gave a new mineral not yet identified.

4. No. 2 glauconite at 300°C. in $MgCl_2$ solution (5 gms. $MgCl_2$ in 20 cc. water) Seven days. The result was a forcing apart of the mica layers and an approach to vermiculite.

In all experiments the air above the solutions was replaced by CO_2.

Ammonium fluoride was added in experiments No. 1 and 2 in the hope of increasing the mobility of the ions. It may have had this effect for the lines in the diagrams of the bomb experiments seem to be slightly sharper. On the other hand, the anomalous line, No. 4, which was mentioned above, becomes rather weak or disappears entirely. This is also true for the specimens heated to 750°C.

Further experiments dealt with the possible replacement of the K ions. If such replacement occurred, shifts of lines might be of value in interpreting the structure. Experiment No. 3 proved a failure in this respect since it formed a new compound. Thallium has been mentioned by Holzner as replacing K in stilpnomelane (15), and also H_2O in stilpnochloran (16). The writer repeated his experiments with essentially similar results. In the case of glauconite considerable amounts of thallium seem to be taken into the structure judging by the increase of the specific gravity of glauconite No. 6 from 2.81 to 3.02 after boiling in thallium nitrate solution for four hours.[2] No appreciable intensity differences were noticed in the powder diagrams of such treated material (see columns 5 and 6, Table 2). Thallium has an atomic number of 81 as compared with 19 for potassium. Its diffractive power, therefore, should be more than four times that of K and when substituted in the structure it should change the intensities markedly. That such a change is not noticeable to the eye is chiefly because at the most only 25 per cent of the K ions were replaced by Tl. The density of the mineral would reach values between 3.9 and 4.0 if the replacement were complete. This makes it probable that only K ions which were at or near the surface of the mineral particles were replaced. It would explain why in biotite (Mora, Minnesota) the density increased from 3.151 to only 3.161 after treatment with thallium nitrate solution. The particle size in the biotite is very much larger and so its surface is only a small fraction of that exposed in glauconite.

[2] After treatment the material was washed very carefully in boiling water several times for long periods.

CHEMICAL CONSIDERATIONS

Many glauconite analyses have been published, some of them excellent, others of doubtful material. For the present discussion only three publications need be mentioned since they contain the latest reliable information. Hallimond (2 and 3) on the basis of twelve selected analyses came to the conclusion that:

> In the groups R_2O_3 and RO the molecular proportions are not constant and do not stand in any simple ratio to the silica and alkalis; the ordinary substitutions of alumina for ferric iron and magnesia for ferrous iron are therefore insufficient to explain the analyses. If, however, the (Fe, Mg)O and $(Fe, Al)_2O_3$ are treated as mutually replaceable, considerable improvement can be brought about.

He gave the formula: $R_2O \cdot 4(R_2O_3, RO) \cdot 10SiO_2 \cdot nH_2O$.

C. S. Ross (1) four years later in a similar study in which seventeen analyses were used, considered glauconite an isomorphous mixture of two end members with the formulas:

$$A = 2H_2O \cdot K_2O \cdot 2(MgO,Fe''O) \cdot 2(Fe_2'''O_3,Al_2O_3) \cdot 10SiO_2 + 3H_2O.$$

$$B = 2H_2O \cdot K_2O \cdot (MgO,Fe''O) \cdot 3(Fe_2'''O_3,Al_2O_3) \cdot 10SiO_2 + 3H_2O.$$

Schneider (4) a year later wrote:

> The composition and variability of glauconite are more accurately expressed by the formula:
>
> $$(K, Na)(Fe, Mg)(Fe, Al)_3Si_6O_{18} \cdot 3H_2O. \ldots$$

The present writer averaged the analyses of each of the three investigators and compared the results. They are remarkably similar. It would be difficult to say which is closest to the truth. They bear out the contention that glauconite is a definite species.

Real difficulty is encountered when one tries to reconcile this glauconite formula with the structure of mica. The structural formula of mica per unit cell is:

$$(OH)_8 \cdot (Mg, Fe'', Fe''', Al)_{8-12}(Si_{12}Al_4)O_{40}.$$

Therefore a biotite hardly ever contains more than 40 per cent SiO_2 while glauconite varies between 46 and 51 per cent SiO_2. The distribution of ions in a typical mica structure is shown diagrammatically in Fig. 1.

Further discussion is best restricted to a glauconite of which we have almost complete data. Glauconite No. 1, whose chemical composition is given in Table 3 is chosen. Its molecular proportions are found in column 1, Table 4. On the assumption that all SiO_2

is contained in the $(Si_3Al)O_{10}$ tetrahedral layers typical of mica the number of SiO_2 "molecules" in a unit cell is twelve. The number of the other oxides is of corresponding size as shown in column 2. Column 3 gives the number of ions. Adjusting the ions now on the

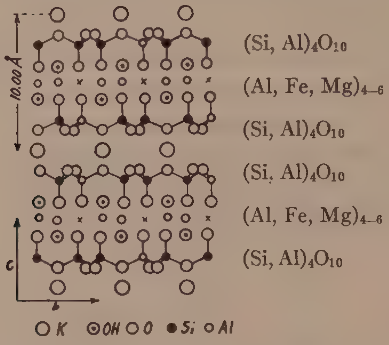

FIG. 1. Diagram showing the sequence of ionic planes in muscovite.

TABLE 3. ANALYSES OF GLAUCONITES AND BIOTITE *X*-RAYED.

	Glauconite No. 1	Glauconite No. 2	Biotite* Mora, Minn.
SiO_2	48.66	49.05	35.67
Al_2O_3	8.46	7.96	14.56
Fe_2O_3	18.80	19.66	3.03
FeO	3.98	0.75	23.23
MgO	3.56	1.17	9.24
CaO	0.62	2.34	1.13
Na_2O	None	0.78	0.49
K_2O	8.31	6.18	8.06
H_2O-	1.94	11.79	0.23
H_2O+	4.62		1.02
TiO_2			3.32
Cr_2O_3			0.03
F			0.16
Total	98.95	99.68	100.22
With CO_2	99.08		

No. 1. Glenn V. Brown, analyst.
No. 2. E. P. Henderson, analyst.
* F. F. Grout, analyst.

basis of the number of K ions present, a typical mica would have in its structure the proportional amounts shown in column 4. This would leave an excess of SiO_2 and H_2O. Is this excess of SiO_2 and H_2O present as submicroscopic opal or adsorbed hydrated SiO_2 around the mica particles? Such SiO_2 would not be easily detectable by *x*-rays. There is a serious objection to such an hypothesis. It could not explain the relative constancy of SiO_2 in glauconites. The optical properties should also be much more variable.

TABLE 4. MOLECULAR AND IONIC RATIOS IN GLAUCONITE NO. 1

	1 Molecular Ratios	2 No. of "Oxide Molecules"	3 No. of Ions		4 Ratio of Ions in Mica	5 No. of Ions in Glauconite	
SiO_2	.8102	12.0	12.0 Si		7.8	14.0 Si	
Al_2O_3	.0830	1.22	2.4	8.2	7.8	2.8 Al	9.5
Fe_2O_3	.1177	1.74	3.5 Fe'''			4.0 Fe'''	
FeO	.0554	0.82	0.8 Fe''			1.0 Fe''	
MgO	.0883	1.31	1.3 Mg			1.5 Mg	
CaO	.0110	0.16	0.2 Ca			0.2 Ca	
K_2O	.0882	1.31	2.6 K		2.6	3.0 K	
H_2O-	.1076	1.59	1.6 H_2O				
H_2O+	.2564	3.80	7.6 OH		5.2	6.1 OH	

The alkali content of many micas is considerably lower than the formulas require. This is also true for glauconites. The deficiency is not serious. It means that not all the K positions between the layers (Fig. 1) need to be occupied. The number of "holes" depends probably upon the number of (Al, Fe''') ions replacing Si, for theoretically for every (Al, Fe''') ion in the $(Si, Al)_4O_{10}$ tetrahedral layer there should be a K ion in the structure. Under the condition under which glauconites form, a deficiency of Al_2O_3 (soluble in some form) is probably the rule, while the supply of SiO_2 in colloidal or other solution is abundant. This would favor a mica structure with a minimum of Al and maximum of Si ions in the $(Si, Al)_4O_{10}$ portions of the layers. On the assumption that of the sixteen (Si, Al) positions in the unit cell only two were occupied by Al instead of four, the Si could be distributed over fourteen positions, and other adjustments made accordingly, as shown in column 5, Table 4. The actual formula for this particular case would be, disregarding excess H_2O:

$$(OH)_8 \cdot K_3(Mg, Fe'', Ca)_{2.7}(Fe''', Al)_{4.8}(Si_{14}Al_2)O_{38-39}$$

while the structural formula would be:

$$(OH)_8 \cdot K_2(Mg, Fe'')_3(Fe''', Al)_6(Si_{14}Al_2)O_{40}.$$

On taking the "molecules" of this actual formula and placing them into the unit cell, a theoretical density of 2.83 is obtained which is slightly less than the one found (2.855). Three H_2O molecules of the analysis are disregarded in this calculation, which is justified by the dehydration curve found by C. S. Ross for this glauconite (1, p. 8).

The Si ions theoretically are confined to the $(Si, Al)_4O_{10}$ tetrahedral layers. But there are reasons to believe that they can and do enter other portions of the structure. While they may not become ions with a covalency of six like Fe''' and Al they probably do replace a few of these occasionally. The best known example of this kind seems to be anauxite (9, p. 87). In kaolinite the ratio $SiO_2:Al_2O_3$ is usually close to 2:1, while in anauxite, which has the same structure, it may be as high as 3:1. It is probable that anauxite, just as glauconite, forms under conditions where the mass action of SiO_2 produced a kaolinite high in SiO_2. There is, however, no avoiding the fact that the unit cell can accommodate at the most fourteen to fifteen Si ions in order to remain below a density of 3.0 (which is probably never exceeded by any glauconite). With fourteen Si ions in the cell, the ratio of Si:(Al, Fe, Mg) is 12:9.5 in column 5, Table 4. This is a trifle too low for the structure proposed which calls for at least 14:10. It is possible that a vacancy or "hole" exists in each unit cell (statistically speaking) which in a mica that formed under conditions of high mobility of ions does not usually exist. That such vacancies may exist in silicates is well known. Due to such deficiency in cations in glauconite OH might easily take the place of O to balance the electrostatic charges. A structural formula covering *all* these proposed changes would be: $(OH)_{6-10} \cdot K_{2-3}(Mg, Fe'', Ca)_{1-3}(Fe''', Al, Si)_{3-6}(Si_{13-14}Al_{2-3})O_{38-40}$

CALCULATIONS OF STRUCTURES

While the powder spectrum photographs of glauconite and mica seem to agree within reasonable limits it was thought best to calculate interplanar distances and intensities for both and compare them with the observed. The muscovite structure C_{2h}^6 as found by Pauling (5) and Jackson and West (6) was used for the calculations. Its corrected coordinates are given by Jackson and West (7). The formula as given by Wyckoff (14, p. 165):

$$I \infty j(A^2+B^2)\frac{1+\cos^2 2\theta}{\sin^2\theta\cos\theta}$$

was used for intensity calculations. The value of the frequency factor j is explained by Wyckoff (14, p. 177). I was divided by 5000 to obtain values of suitable size. The ionic numbers of the atoms were used for their scattering power F. O and OH $=10$, Si and Al $=10$. The Fe″, Fe‴, Al and Mg combinations which occupy the Al positions were assumed in such proportions to give each a scattering power of 20. It was assumed that all K positions are occupied. K $=18$. Any reduction in their number would materially influence the theoretical intensities as indicated in column A, Table 5 by the + and − signs for the first twenty planes. Column B gives the calculated intensities as they would be if Fe‴ occupied the Al positions in the (Si_3Al) tetrahedral layers and all of the Al ions were in the positions between the tetrahedral layers. For this case the assumption was also made that only two-thirds of the K positions are occupied. This does not mean that all Al is actually replaced by Fe‴ in the positions of the (Si_3Al)O_{10} tetrahedral layers. It is simply a possibility. Values somewhere between *A* and *B* are most likely the best.

The intensities for biotite, Table 5, were calculated as those of column *A*, Table 5, except that the additional four Al positions of the structure which are vacant in muscovite were also filled with ions of scattering power, $F=20$.

Considering the nature of the formula used, the agreement between observed and calculated intensities is fair for most planes. The *hhl* and *0kl* reflections do not agree any better than for other layer structures in which this behavior was pointed out previously (11, p. 417).

The indices of two or three weak reflections were not calculated on account of the very large amount of labor involved in comparison to the benefits that could be derived. On the other hand, several dozen calculated reflections of low intensity were omitted from Table 5 on account of lack of space. The rate of decrease in observed intensities in the layer structures whose particles are

Table 5. Theoretical and Observed Intensities for Unit Cells of Glauconite and Biotite

Glauconite					Biotite from Mora, Minn.		
Observed Int.	Theoretical Intensity		d	Indices	d	Theoretical Intensity	Observed Int.
	A	B					
1–2	190+	160	9.970	002	10.015	390	4
	0	27	4.985	004	5.007	5	
2	3−	2	4.537	020	4.616	3	½
	21+	39	4.522	110	4.601	21	
	52+	56	4.484	$11\bar{1}$	4.562	4	
	8+	24	4.422	021	4.497	8	
	14−	4	4.341	111	4.408	1	
	0	2	4.241	$11\bar{2}$	4.308	0	
	14+	14	4.128	022	4.192	1	
	3	0	4.007	112	4.065	3	
1–2*b*	48−	31	3.878	$11\bar{3}$	3.932	6	
	32−	27	3.746	023	3.797	32	
	0+	0	3.615	113	3.661	16	
	52−	48	3.484	$11\bar{4}$	3.526	52	
2–3	70−	42	3.349	024	3.394	22	
	72+	77	3.323	006	3.338	111	5
1*b*	64−	62	3.231	114	3.266	64	½
	2−	4	3.111	$11\bar{5}$	3.143	4	
	59−	55	2.996	025	3.026	59	
Ind.	51−	40	2.886	115	2.913	16	
	42−	36	2.781	$11\bar{6}$	2.807	42	
	0	0	2.682	026	2.705	10	
	14	23	2.608	200	2.654	5	
	24	26	2.608	$13\bar{1}$	2.654	8	
3–4	25	19	2.587	116	2.609	25	
	34	26	2.579	$20\bar{2}$	2.624	53	2
	78	62	2.579	131	2.624	118	
	2	1	2.492	008	2.503	8	1
	2	9	2.471	202	2.512	0	
	5	20	2.471	$13\bar{3}$	2.512	0	
2	60	62	2.399	$20\bar{4}$	2.436	83	2
	101	119	2.399	133	2.436	142	
½	17	11	2.268	040	2.307	6	
	34	23	2.266	$22\bar{1}$	2.306	34	½*b*
	1	1	2.261	220	2.301	2	
	31	21	2.253	041	2.292	31	
	27	18	2.242	$22\bar{2}$	2.281	8	
	0	0	2.233	204	2.264	0	
	0	0	2.233	$13\bar{5}$	2.264	2	
	26	18	2.228	221	2.266	26	

TABLE 5. CONTINUED.

Glauconite					Biotite from Mora, Minn.		
Observed Int.	Theoretical Intensity		d	Indices	d	Theoretical Intensity	Observed Int.
	A	B					
	0	0	2.211	042	2.249	4	
	19	13	2.170	222	2.206	5	
1*b*	6	4	2.146	043	2.181	6	
	70	45	2.145	$20\bar{6}$	2.173	90	1–2
	133	85	2.145	135	2.173	173	
	5	7	2.064	044	2.096	0	
1*b*	46	53	1.994	0010	2.003	33	2–3
	20	12	1.971	206	1.994	30	
	40	27	1.971	$13\bar{7}$	1.994	61	
	7	4	1.886	$20\bar{8}$	1.908	12	½
	13	9	1.886	137	1.908	25	
	0	0	1.729	208	1.746	1	
	0	0	1.729	$13\bar{9}$	1.746	2	
2*b*	11	4	1.662	0012	1.669	17	3
	40	37	1.656	$20\overline{10}$	1.672	52	
	78	71	1.656	139	1.672	101	
	18	15	1.621	313	1.647	7	
	13	10	1.621	$31\bar{5}$	1.647	4	
	13	12	1.575	314	1.599	13	
4	29	23	1.522	2010	1.535	38	4
	51	42	1.522	$13\overline{11}$	1.535	68	
	50	47	1.512	060	1.538	60	
	104	94	1.512	$33\bar{1}$	1.538	121	
½–1	28	27	1.495	$33\bar{3}$	1.520	41	0.5
	24	21	1.495	331	1.520	35	
	4	5	1.461	$20\overline{12}$	1.473	8	½
	6	6	1.461	1311	1.473	12	
	10	2	1.447	$33\bar{5}$	1.471	18	
	7	2	1.447	333	1.471	13	
	5	6	1.424	0014	1.430	7	1
½	1	1	1.376	$33\bar{7}$	1.397	0	
	1	1	1.376	335	1.397	0	
b Ind.	41	46	1.350	2012	1.360	50	2–3
	83	93	1.350	$13\overline{13}$	1.360	105	
2	23	27	1.308	$40\bar{2}$	1.331	17	½–1
	58	70	1.308	260	1.331	44	
	41	37	1.304	400	1.327	50	
	59	56	1.304	$26\bar{2}$	1.327	75	
	3	8	1.299	$20\overline{14}$	1.309	1	
	8	10	1.299	1313	1.309	4	

TABLE 5. CONTINUED.

Glauconite					Biotite from Mora, Minn.		
Observed Int.	Theoretical Intensity		d	Indices	d	Theoretical Intensity	Observed Int.
	A	B					
	9	10	1.293	$33\bar{9}$	1.311	16	½
	9	7	1.293	337	1.311	16	
	21	14	1.290	$40\bar{4}$	1.312	27	
	0	1	1.290	262	1.312	1	
½	18	20	1.279	402	1.301	24	½
	28	55	1.279	$26\bar{4}$	1.301	63	
1	8	10	1.252	$40\bar{6}$	1.272	12	½
	14	19	1.252	264	1.272	23	
	31	26	1.246	0016	1.251	43	1
	20	18	1.236	404	1.256	27	
	4	0	1.236	$26\bar{6}$	1.256	9	
½	13	7	1.207	2014	1.216	18	½
	31	20	1.207	$13\overline{15}$	1.216	43	
	11	15	1.205	339	1.220	5	
	0	2	1.205	$33\overline{11}$	1.220	1	
	0	0	1.199	$40\bar{8}$	1.218	1	
	1	2	1.199	266	1.218	3	
	c_0		20.03		20.11		
	b_0		9.07		9.23		
	a_0		5.24		5.33		
	β		95° 00′		95° 04′		

b = unusually broad line.
Ind. = indistinct.

below their optimum size as in glauconite shows a noteworthy behavior. As the third index of planes increases—in other words, as the angles of the planes which they make with the basal plane become relatively small—the decrease in intensity is considerably greater than expected from the formula given. Therefore, "steep" planes reflect relatively more intensely. The calculated sizes of the unit cells and β angles, Table 5, of biotite and glauconite agree quite well with those observed. The mean value of β for all glauconites is close to 95°.

CONCLUSIONS

It is shown by powder spectrum photographs that the structures of glauconite and mica are almost identical. The constants of the minerals, choosing a biotite for comparison, are:

	Glauconite) (Avge. of 6 specimens)	Biotite (Mora, Minnesota)
c_0	20.03 Å	20.11 Å
b_0	9.07	9.23
a_0	5.24	5.33
β angle	95°00′	95°04′
Axial ratio =	.5773:1.000:2.208	.5773:1.000:2.179

The theoretical plane 100 for a β angle of 90° as usually quoted for biotite becomes the plane $30\bar{2}$ in this new orientation, which corresponds to that of muscovite.

Glauconite is as stable as biotite at temperatures as high as 750°C., which is additional proof of its structure. Glauconite absorbs considerable amounts of thallium ions which probably replace K ions to an extent not exceeding 25 per cent.

The high SiO_2 and H_2O content of glauconite seems to be due to the environment in which it is formed. Excess of soluble SiO_2 over available soluble Al_2O_3 gives rise to a higher Si:Al ratio in the $(Si, Al)_4O_{10}$ tetrahedral layers than in mica. It may even cause substitution of a few Si ions for Fe‴ or Al in approximately the positions having a covalency of 6. These Si ions would not necessarily be hexavalent, however.

It is also thought that in comparison with a muscovite glauconite may have occasional vacant positions or "holes" in its structure. A formula which would take into account such probabilities is: $(OH)_{6-10}\cdot K_{2-3}(Mg, Fe'', Ca)_{1-3}(Fe''', Al, Si)_{3-6}(Si_{13-14}Al_{2-3})O_{38-40}$. The formula for a specific case (glauconite No. 1) is:

$$(OH)_{6-8}\cdot K_3(Mg, Fe'', Ca)_{2.7}(Fe''', Al, Si)_{4.8}(Si_{13}Al_3)O_{38-39}.$$

Calcium probably would be permissible to only a fraction of a per cent in glauconite. Of the H_2O given in analyses, probably about half may be adsorbed.

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PRIMARY NATIVE SILVER ORES AT BATOPILAS, MEXICO, AND BULLARD'S PEAK, NEW MEXICO

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Introduction

Many of the silver deposits on which there is seemingly general agreement as to their hypogene (primary) origin, and particularly those deposits in which native silver is a fairly common constituent, show a characteristic association of nickel and cobalt arsenides with native silver or other silver minerals. In view of the generally recognized hypogene origin of the nickel and cobalt arsenides this association has been used as one of the criteria for the primary origin of native silver. Such an association has been observed in a number of localities and described by various writers. It is one of the striking features of the silver deposits in the Erzgebirge of Europe. A similar association in the mines of the Cobalt district of Ontario is so well known that further mention is not necessary.

In addition to these two best known of the districts where this association is common, Bastin[1] has also described silver deposits near Wickenburg, Arizona, where the same association is found and where he believes much of the native silver as well as other silver minerals are due to hypogene mineralization. More recently the deposits of the Great Bear Lake district have been studied by Kidd[2] where a similar association is noted. Piepoli[3] has described the ores of Sarrabus, Sardinia, where cobalt and nickel arsenides are associated with native silver and pyrargyrite. The silver minerals at this locality are also considered to be of primary origin.

At Sabinal, in Chihuahua, Mexico,[4] native silver and other silver minerals are intimately intergrown and associated with similar arsenides and with galena, sphalerite, chalcopyrite and pyrite.

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[1] Bastin, E. S., Primary Native Silver Ores near Wickenburg, Arizona, and their Bearing on the Genesis of the Silver Ores of Cobalt, Ontario: Bull. **735,** *U. S. Geol. Survey*, 1922, pp. 131–154.

[2] Kidd, D. F., A Pitchblende Silver Deposit, Great Bear Lake, Canada: *Econ. Geol.*, vol. **27,** pp. 149–159, 1932.

[3] Piepoli, M. P., Etude microscopique de quelques minerais des filons argentiferes du Sarrabus (Sardaigne): *Soc. Franc. Miner.*, Bull. **56,** (6–7–8), pp. 277–302, 1933.

[4] Krieger, P., Primary Silver Mineralization at Sabinal, Chihuahua, Mexico: *Econ. Geol.*, vol. **30,** pp. 242–259, 1935.

Such an association of minerals and the manner of their occurrence leaves little doubt as to the hypogene origin of the native silver at this locality also.

Recent study of material from two additional localities shows a similar mineralogical association. At Batopilas, in Chihuahua, Mexico, silver mineralization occurs which appears to be typically primary, with native silver and the sulpho-salts of silver intimately intergrown with arsenides of cobalt and nickel, arsenopyrite, and with the sulphides of zinc and lead. The field relations at Batopilas and the fact that some of the veins have been mined to depths of 1700 feet below the surface outcrops also support the view that silver mineralization here is of hypogene origin.

A specimen of native silver ore from the Bullard's Peak district, New Mexico, also shows native silver associated with a cobalt-nickel arsenide. Although this locality was not examined in the field, a microscopic study of the ore indicates that the native silver is probably also of a primary nature.

Batopilas District, Chihuahua, Mexico

This district is located in one of the more rugged portions of the Sierra Madre mountains of southwestern Chihuahua, approximately 60 miles south of Creel, the end of a branch line of the Mexican Northwestern Railroad. The mines have been operated intermittently since their discovery in 1632 and have yielded, at times, remarkably rich bonanzas of native silver ore. Despite the fact that Batopilas has been known as a silver producing district for more than three centuries and has yielded over 50 million ounces of silver, surprisingly little has ever been published concerning the geology or mineralization of this famous locality. The mines have long been noted for the beauty and variety of native silver specimens and crystals of ruby silver, examples of which can be found in many of the larger museums throughout this country and in Mexico.

The ore occurs in a series of well defined fissure veins in diorite or andesite that has been intruded by a fairly coarse textured granite. By far the greater portion of the veins, particularly those from which the bulk of the silver has been produced, strike nearly north-south. In the San Miguel mine a few minor cross veins with a nearly east-west strike have also yielded considerable silver. The veins vary in dip from 65 degrees west to vertical. Gangue material in the veins consists almost entirely of calcite with only minor amounts of quartz and barite. A preliminary study of the geology of the district indicates that the mineralization was probably connected with the granite intrusive.

The present study is intended to illustrate some of the microscopic features of the ore minerals and to show the association of silver minerals with cobalt and nickel arsenides and with the sulphides of other metals. A more detailed study of the general geologic features of the district is being planned.

THE ORES. Most of the ore derived from the veins consists of native silver with subordinate amounts of polybasite, argentite, proustite and pyrargyrite. Pyrite, galena and a very light, amber colored sphalerite are also present. Large masses of cerargyrite are also reported to have been mined near the surface outcrop of the veins during early mining operations.

Microscopic study of the ore shows additional minerals which, from their association and structural relations, afford good evidence for the primary origin of these ores and support the field evidence for such an origin. Although safflorite and rammelsbergite, together with arsenopyrite, rarely occur in sufficient quantity as to be visible to the unaided eye, they were readily observed in nearly all of the polished sections examined under the microscope. These minerals are considered significant as regards the origin of the silver minerals. Rhombic forms and stellar shaped twin crystals of safflorite and rammelsbergite are fairly common and are intimately associated with native silver, galena, sphalerite and the sulpho-salts of silver. Some of these are shown in Figs. 1 to 4. Euhedral crystals of arsenopyrite are likewise common and are also associated with native silver.

A slight time interval may have existed between deposition of the various minerals in this group but they are so intimately intergrown with each other and present such closely related structural and textural features that a similar origin must be assigned to each of them, including the silver minerals. Thus, the ores at Batopilas resemble, in many respects, other silver deposits where similar mineral assemblages have been noted and where native silver is generally recognized as being of primary origin. The absence of evidence indicating oxidation or enrichment by supergene processes supports this conclusion.

An exact order of deposition for each of the metallic minerals is difficult to establish. In general, however, the silver minerals belong to the last stages of metallic mineralization and appear to be, in greater part, later than the arsenides and arsenopyrite. Evidence for this is to be seen by the manner in which native silver has penetrated between individual crystals of these minerals and by modification of their crystal form when in contact with native silver. Native silver is distinctly younger than sphalerite. This is

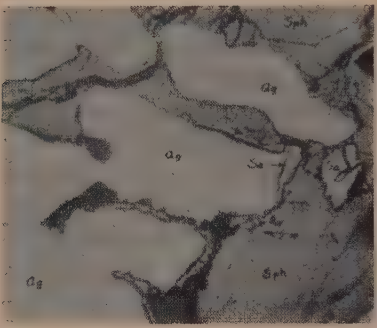

FIG. 1. Native silver (Ag) associated with sphalerite (Sph) and safflorite (Sa). Batopilas, Chihuahua, Mexico. ×75.

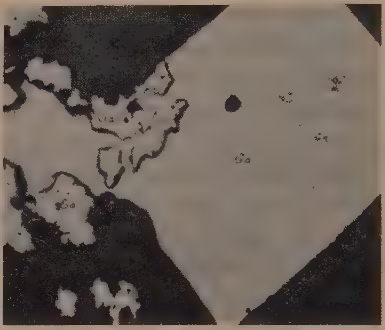

FIG. 2. Groups of rhombic and twin crystals of safflorite (Sa) associated with galena (Ga) and native silver (Ag). Batopilas, Chihuahua, Mexico. ×75.

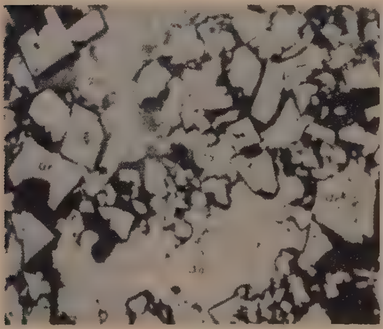

FIG. 3. Showing native silver (Ag) penetrating between euhedral crystals of arsenopyrite (Ar). Some of the arsenopyrite appears to have been partially replaced by native silver. Batopilas, Chihuahua, Mexico. ×75.

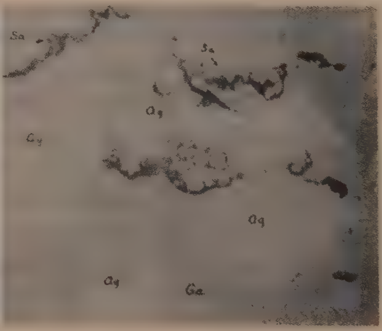

FIG. 4. Stellar shaped twin crystals of safflorite (Sa) associated with native silver (Ag) and galena (Ga). Batopilas, Chihuahua, Mexico. ×75.

indicated by small veinlets of native silver within fractures in sphalerite. Likewise, native silver appears to be, at least in part, later than galena, as indicated by small hybrid veinlets which sometimes penetrate the latter mineral. Such veinlets, however, are not abundant and in most of the specimens examined the relation between these two minerals could not be definitely established. It is quite possible that contemporaneous deposition of galena and native silver took place, with native silver continuing to be deposited over a slightly longer period of time.

The relations observed between the sulphosalts of silver also indicate contemporaneous deposition. Specimens available did not show these minerals in contact with native silver so this relationship could not be determined. Remnant crystals of the arsenides within polybasite and proustite indicate that the arsenides were deposited earlier in the sequence of mineralization.

The occurrence of well formed, stellar shaped twins and euhedral rhombic crystals of safflorite and rammelsbergite within galena and sphalerite suggests that the arsenides are the younger of these minerals. Pyrite is found largely in the wall rock of the veins and has not been observed in contact with other metallics. Under normal conditions it would be expected to be one of the earlier metallic minerals to crystallize.

Although an exact sequence in the deposition of these minerals cannot be established the general order in which they were deposited appears to be as follows: (1) deposition of the sulphides, including pyrite, sphalerite and galena; (2) deposition of the nickel and cobalt arsenides and arsenopyrite; (3) deposition of the silver minerals, with native silver probably continuing as one of the last of the metallics.

BULLARD'S PEAK DISTRICT, NEW MEXICO

Waller and Moses,[5] in 1892, described what they thought was probably a new nickel arsenide from this locality. In addition to arsenic, analysis showed the presence of nickel, cobalt and iron. The combined ratio of these elements corresponded to the ratio for skutterudite, the cobalt tri-arsenide, and the name "nickel-skutterudite" was suggested for the mineral.

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[5] Waller, E., and Moses, A. J., A Probably New Nickel Arsenide: *School of Mines Quarterly*, vol. **14**, pp. 49–51, 1892–3.

Recent x-ray work by Holmes[6] on a specimen of this material in the Egleston Collection at Columbia University shows a diffraction pattern corresponding to that of skutterudite from Skutterud, Norway, confirming the earlier work of Waller and Moses.

The nickel-skutterudite is found intimately associated with native silver in a gangue of quartz, calcite and siderite. According to early accounts of its occurrence, the arsenide was not present in the upper 100 feet of the mine. Below the 100 foot level, however, and to a depth of 690 feet, the lowest workings at that time, it was found in appreciable quantities.

Polished surfaces of the material show a striking similarity to some of the ores of Cobalt, Ontario, and to some of the European deposits. The material also contains a variety of structures which may have a bearing on the paragenesis of the two minerals and appear to support the recent experiments of Schouten[7] on structures obtained by synthetic replacement of similar materials.

The native silver occurs in arborescent masses, most of which show good crystalline form and is nearly always surrounded by bands of nickel-skutterudite. In addition to the bands surrounding native silver, the nickel-skutterudite also occurs as groups of crystals similar to the tubercle forms described by Bastin[8] from the Cobalt district. In such cases the arsenide surrounds gangue minerals with the interior of the crystal groups entirely devoid of native silver or but partially filled with native silver. In most instances, however, the interiors of the crystal groups are entirely filled with native silver, many of which show good crystal faces against the nickel-skutterudite.

One conclusion regarding the paragenesis of the two minerals appears evident. Some of the native silver was deposited after deposition of the arsenide. This is shown by fractures and small veinlets in the arsenide that contain native silver. In some instances the opposite walls of the fractures match perfectly so that simple fracture filling appears to have taken place. Other instances, however, show irregular veinlets with walls that do not match, indicating that some replacement may also have occurred.

[6] Holmes, R. J., Paper presented before the 15th annual meeting of the Mineralogical Society of America at Rochester, New York, December, 1934.

[7] Schouten, C., Structures and Textures of Synthetic Replacements: *Econ. Geol.*, vol. **29**, pp. 611–658, 1934.

[8] Bastin, E. S., Primary Native Silver Ores of South Lorraine and Cobalt, Ontario: *Econ. Geol.*, vol. **20**, pp. 1–24, 1925.

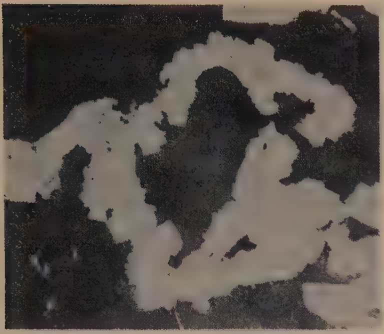

FIG. 5. One of the tubercle-like forms of nickel-skutterudite with the interior entirely devoid of native silver. Note the development of crystal faces on both the inner and outer borders. Bullard's Peak district, New Mexico. ×75.

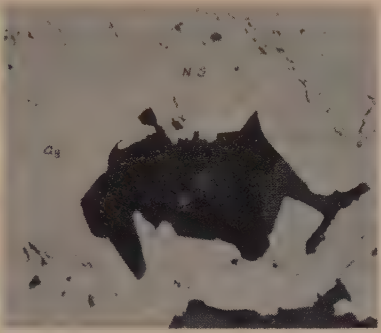

FIG. 6. Another of the tubercle-like forms of nickel-skutterudite (N–S) partially filled with native silver (Ag). Where the two minerals are in contact with each other the crystal faces of the nickel-skutterudite appear to have been modified, suggesting that the interior of these forms were filled with native silver and that some replacement of the arsenide was accomplished. Bullard's Peak district, New Mexico. ×75.

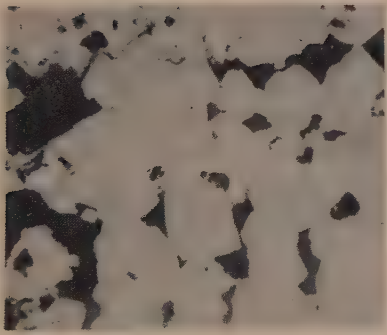

FIG. 7. Dendritic crystals of native silver within nickel-skutterudite. Bullard's Peak district, New Mexico. ×75.

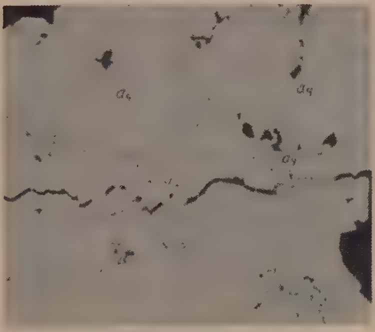

FIG. 8. Fracture in nickel-skutterudite filled with native silver (Ag) and gangue. Dendritic crystals of native silver are also present in the arsenide. Bullard's Peak district, New Mexico. ×75.

Several features concerning the tubercle-like forms of nickel-skutterudite are of interest because they show evidence which suggests that native silver may have been deposited by so-called "centrifugal" replacement as suggested by Schouten[9] as the result of his work on synthetic structures. Where these forms surround only gangue minerals they show a striking development of crystal faces on both the inner and outer border of the arsenide. Such a form is shown in Fig. 5. When the interiors of these tubercle-like forms are filled with native silver, however, the crystal faces of the arsenide on the inner border are largely destroyed and the crystal form of the native silver is substituted.

Several instances were noted where a tubercle was but partially filled with native silver. Where the two minerals are in contact with each other the crystal faces of the arsenide appear to have been destroyed, but where they are not in contact, the crystal faces of the arsenide are still present. Occasional remnants of the arsenide within native silver are also present.

In presenting the above evidence the author fully realizes that it is only suggestive and cannot be taken as conclusive evidence in itself, since it is quite possible that simultaneous deposition of the two minerals may have taken place to a certain extent. However, in view of the fact that at least some of the native silver is later than the arsenide, the weight of the evidence presented supports the younger age of the native silver.

The question as to the sequence of the two minerals does not detract from the significance of this association as regards the origin of the native silver. This association of arsenides with native silver appears to be so typical of those native silver ores which are generally recognized as being due to hypogene mineralization that a similar origin for this ore seems probable. This conclusion is further supported by the fact that oxidation products are absent in the ore.

Conclusions

The association of nickel and cobalt arsenides with native silver is becoming increasingly common as silver deposits are being studied in greater detail. Two additional localities showing this association are presented in the present study.

At Batopilas, native silver is associated with safflorite and rammelsbergite, and with arsenopyrite, galena and sphalerite. Such an assemblage of minerals and the manner of their occurrence indicates that native silver at this locality is of hypogene (primary) origin. A preliminary study of the field relations supports the microscopic evidence for such an origin. So far as known, this is the second locality in Mexico where such an association has been reported and where native silver is believed to be of primary origin.

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[9] Op. cit.

Microscopic study of material from the Bullard's Peak district, New Mexico, shows native silver associated with a nickel-skutterudite. This association is strikingly similar to the association of native silver and arsenides in the Cobalt district of Ontario. A primary origin for the native silver at this locality is also advocated. The material from Bullard's Peak district also shows evidence suggestive of centrifugal replacement of the arsenide by native silver.

The trip into Batopilas was made in the company of Mr. J. Conness Shepard, of Chihuahua City, to whom sincere appreciation is extended for information concerning the mines during their operation by Governor Alexander Shepard.

MONAZITE FROM WEST PORTLAND TOWNSHIP QUEBEC

Hugh S. Spence[1] and O. B. Muench[2]

Part i. Hugh S. Spence

Introductory

In 1929, during a survey of feldspar mining operations, a small quarry on lot 2, range V of West Portland Township, in the Lièvre river section of Quebec—Lat. 45°45′30″N, 75°36′30″W (N side of river)—was visited. The property was idle, and the one small pit full of water; pegmatite was exposed only immediately around the pit and for about 75 feet up a low cliff-face rising abruptly from it. Examination of the cliff section, from which some rock had been scaled down during mining operations, disclosed nothing of interest in the way of unusual accessory minerals, the dyke-mass being the usual mixture of pink microcline (perthite) and massive white quartz, typical of the general run of Canadian pegmatites in both Quebec and Ontario. About 800 tons of feldspar were reported to have been shipped from the property.

On examining the waste-dump, however, some interesting minerals were found. A single large block of rock, lying on the top of the dump and evidently one of the last pieces raised from the pit, was found to contain considerable euxenite. In addition to the euxenite, and usually in close association with it, were found a number of crystals of the monazite described in this paper. The block was broken up, and about 5 pounds of clean monazite was recovered, in the form of rough crystals and cleavage pieces. The largest piece measured about 6″×3″×1″ and weighed ¾ pound. Probably an equal amount of the mineral was secured as matrix specimens, mixed with euxenite, black tourmaline and feldspar. Further search in the dump, on this occasion and on several subsequent visits, was rewarded by the finding of more material, most of it as clean cleavage pieces up to ½ pound in weight. The total collected was about 20 pounds.

The discovery was of interest, since monazite is not a common mineral in Canadian pegmatites and has been recorded from only a few dykes. This West Portland occurrence was briefly noted in a previous paper[3] by the writer. The only other recorded occurrence of the mineral up to the time that paper was written was at the old Villeneuve mine, in Villeneuve Township, Que., where a nodular mass of 12 pounds weight is stated[4] to have been found many years ago. In a later paper,[5] Ellsworth has described monazite from a pegmatite in Dickens Township, Nipissing District, Ont., and Walker mentions[6] its occurrence at the Huron claim, on the Winnipeg river, near Pointe du Bois, Man. The writer has observed it in small amount, but only as minute anhedral grains, in the beryl pegmatite on lots 23 and 30, concession XV of Lyndoch Township, Renfrew County, Ont., and also in the syenite pegmatite of Cardiff Township, Haliburton County, Ont.

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[1] Mines Branch, Department of Mines of Canada. Submitted with the permission of the Director, Mines Branch.

[2] New Mexico Normal University, Las Vegas, New Mexico.

Since a sufficient amount of the West Portland material was collected to serve research purposes, the writer sent a quantity of it to Dr. A. C. Lane, suggesting that it might be useful for age determination purposes, and also as a source of thorium-lead, in connection with the work of his Committee on the Measurement of Geologic Time. In this, Dr. Lane concurred, and a portion of the material was turned over to the second author for detailed chemical analysis, the results of which are given in Part II of this paper. Dr. Lane also sent a portion to Dr. Hecht, in Vienna, for micro-analysis: the results of this are likewise appended.

DESCRIPTION OF THE PEGMATITE

The West Portland pegmatite is a small dyke, not over 50 feet across at its widest surface exposure. This exposure occurs at the base of a low cliff-face in the north side of a small, open valley trending westward through the hills running north and south along the west side of the main Lièvre river valley. The single small pit opened on the dyke is situated at the foot of the cliff, and is about 20 feet square and 25 feet deep. Having sheer walls, and being half full of water, no examination of the opening was possible. The dyke follows an erratic, diagonal course up the cliff from the pit, pinching out upwards until, at the top, it becomes a mere stringer. Beyond the pit, in the opposite direction, it is covered by waste and its further course is hidden beneath soil and drift cover. The country rock is biotite gneiss.

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[3] Pegmatite Minerals of Ontario and Quebec, *Am. Mineral.*, vol. **15**, p. 448, 1930.

[4] *Geol. Surv. Can.*, *Ann. Rep.*, vol. **2**, 1886, pp. 10–11T.

[5] Monazite Coloured by Carbon from Dickens Township, Nipissing District, Ontario: *Am. Mineral.*, vol. **17**, pp. 18–28, 1932.

[6] Rare Minerals in Pegmatite, Pointe du Bois, Manitoba, *University of Toronto Studies*, no. **30**, Contributions to Canadian Mineralogy, p. 12, 1931.

As stated above, the accessible portions in the cliff-face disclosed nothing of interest and no unusual minerals were noticed. It would appear that most, if not all, of the euxenite, monazite and other minerals noted below, which were found on the dump, came from the lower portion of the pit, since the material secured was mostly found in the upper layer of waste. The operator, also, stated that work had been discontinued on account of the increasing amounts of impurities encountered with depth.

Below are listed the various minerals found, in order of their relative abundance.

MINERALS PRESENT IN THE PEGMATITE

Feldspar.—The rock of the dyke is predominantly a pale, pink perthite, intergrown with irregular zones of massive, white quartz.

In the zones carrying the monazite and euxenite, the feldspar is usually a grey or flesh-coloured albite, often showing the colour-play of peristerite. Sometimes, this albite shows the dark-reddish discoloration often noticed around radioactive minerals.

Quartz.—Only the normal massive, white type, common to pegmatites, occurs.

Tourmaline.—Black schorl is abundant, as fine needles penetrating feldspar, but more commonly as stout prisms, up to 2″ across and 12″ long. It is particularly abundant in the rock containing the euxenite and monazite. Having come from near the surface, most of the tourmaline is weathered and friable.

Euxenite.—Occurs as platy crystals up to 2″ across, but being firmly embedded in the rock, most of the material secured was fragmentary. The mineral is fresh and lustrous, and has, variously, a granular or a smooth, conchoidal fracture, the former being most common. On the edges, pieces show reddish-brown translucency. No analysis of the mineral was made.

Monazite.—Monazite and euxenite are present in about equal amount, and appear to be usually closely associated. Frequently, crystals of both occurred side by side, often accompanied by tourmaline. The monazite is uniformly of a dark-reddish colour. Much of the material found was more or less severely weathered, possessed little lustre and was of a dark, brick-red shade. A number of cleavage pieces were secured, however, which were relatively fresh. Under the binocular microscope, most of the mineral was found to show irregularly distributed zones or grains of a yellowish, resinous substance, which may be residues of fresh monazite that have escaped oxidation or, perhaps, altered fergusonite (see below). The same feature was observed on examining a thin section. In the feldspar enclosing the monazite, small anhedral grains of a similar yellow, resinous mineral are often present: whether these are monazite or fergusonite was not determined. Many of the more highly altered crystals also were found to contain numerous microscopic grains of feldspar and quartz, and often flakes of muscovite. Only the freshest monazite was selected for the purpose of the analysis given below.

Allanite (?).—A few specimens of a severely altered, greenish-black mineral that may originally have been allanite were found: the material was too highly weathered to be identified with certainty.

Titanite—Ilmenite: mixture.—A number of pieces, up to ½ pound in weight, of a heavy, greyish-black mineral, apparently having obscure crystal outline and showing indistinct cleavage, were collected. The material was nearly all severely altered (kaolinized) and much of it was of a clayey or earthy consistency, though the freshest specimens showed a matt, sub-metallic appearance, with a medium-granular texture. Alteration had proceeded too far to permit of a precise determination, but the original material was probably titanite. The ground-mass contains scattered grains of a heavy, black, magnetic mineral, which is probably secondary ilmenite. Some calcite, also, is present, and a little pyrite.[7] Analysis showed:

SiO_2	8.80
Al_2O_3	3.97
Fe_2O_3	55.83
TiO_2	22.00
CaO	8.80
	99.40

Fergusonite (?).—In the albite enclosing the monazite and euxenite, small amounts of a mineral having the macroscopic characteristics of fergusonite were observed. It occurs mostly in small, anhedral grains, seldom as much as 5 mm. across, but occasionally exhibits an approach to square prismatic form. When fresh, it is black and has vitreous lustre, but it evidently is readily altered, most of the larger pieces consisting of black cores surrounded by a brown zone which, in turn, is bordered externally by progressively lighter-coloured substance, the outer shell being pale yellow. Such altered material is translucent to transparent, possesses resinous lustre and exhibits conchoidal fracture similar to that of the fresh cores.

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[7] The writer is indebted to H. Berman for the mineral determinations given.

Cyrtolite.—Small crystals of pale grey cyrtolite are fairly frequent in the finer-textured pegmatite enclosing the monazite-euxenite association. Usually single individuals, they range from 3 mm. in diameter down.

Muscovite.—Very little mica is present in the pegmatite, all of it muscovite of a pale, yellow shade. This occurs as occasional small aggregates of flakes, seldom over $\frac{1}{2}''$ across, and also as thin flakes coating tourmaline.

Tengerite.—On fractures (in part cleavage partings) in the albite enclosing the monazite, thin, radiated, dendritic crusts of a white secondary mineral with pearly lustre were noted. Specimens of this have been determined by Dr. Foshag as tengerite (yttrium carbonate). This is the first recorded occurrence of this mineral in Canada.

Specularite.—Small cavities in corroded feldspar are sometimes encrusted with minute crystals of specularite.

While none of the "accessory" minerals listed above were seen in place, the character of the rock enclosing them suggests that they are localized within a finer-grained, albite phase of the pegmatite, of pocket or "schlieren" type, and of limited extent. Just where this zone occurs in the dyke-mass could not be ascertained definitely, but there was some indication that it might be located in the east wall of the pit and near the contact of the pegmatite with the enclosing gneiss.

PART II.

THE ANALYTICAL DETERMINATION OF THORIUM, LEAD AND URANIUM IN THE WEST PORTLAND MONAZITE

O. B. MUENCH

Thorium bearing minerals free from uranium and ordinary lead are of rare occurrence, especially in such quantities that an atomic weight determination can be made of the lead. A specimen of the monazite described in part I of this paper was sent to Dr. A. C. Lane, hoping that it might be nearly, if not quite free from uranium and at the same time high enough in lead and thorium to make it a practical source of these two elements.

An age determination of this mineral by the lead-uranium method was undertaken by the author so that the above information could be obtained and at the same time furnish some data on the age of the mineral and the pegmatite.

The sample as sent by the author of part I of this paper, was purified at Harvard University by treatment with heavy liquids. This sample (80 grams), as it reached the Laboratory of the University of Missouri, was ground in an agate mortar to pass an 80 mesh sieve. The sieve had no exposed solder on it to contaminate the mineral.

There was not enough of the sample available to do any extensive experimentation with different methods of analysis. Fenner[8] gives a detailed method for the analysis of monazite for thorium, lead and uranium, such as is used in age determinations.

Fenner's method was followed with but few modifications. The method consists essentially of a precipitation of the thorium together with the rare earths by means of oxalic acid, then thorough drying and conversion of the oxalates into nitrates by means of fuming nitric acid. This is followed by repeated precipitation with hydrogen peroxide (30%) in a slightly acid or neutral solution containing a strong electrolyte. This precipitate after conversion to the oxalate is ignited and weighed as ThO_2.

Instead of three precipitations with perhydrol, four were found necessary in order to get a thorium precipitate which, from its white color, showed no evidence of cerium.

The lead determination was made as carefully as possible on 30 gram samples, again following Fenner's method. Tested reagents and those known to be lead free were used throughout this analysis. The final precipitate was purified as suggested by Fenner. Care was taken not to heat this lead sulfate above 500°–600°C., for, according to Hillebrand and Lundell,[9] the decomposition of lead sulfate begins at 637°C. and becomes energetic at 705°C.

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[8] Fenner, C. N., The Analytical Determination of Uranium, Thorium and Lead as a Basis for Age Calculations: *Am. J. Sci.*, p. 369, Nov. 1928.

[9] Hillebrand and Lundell, *Applied Inorganic Analyses*, p. 545.

The solutions from the lead determinations were taken for the determination of uranium, and although certain evidence of the presence of uranium was obtained, the results were not quantitative. Because of the comparatively small amount of mineral available, the quantitative results for the uranium were obtained by the emanation method on two gram samples.[10]

EXPERIMENTAL RESULTS

Thorium

Analysis	Wt. of sample (grams)	Wt. of ThO_2 (grams)	Per cent Thorium
1	2.8045	0.1082	3.39
2	2.7029	0.1040	3.38
3	2.5007	0.1010	3.55
			Average = 3.44

Lead

Analysis	Wt. of sample (grams)	Wt. of $PbSO_4$ (grams)	Per cent Lead
1	30.0000	0.0317	0.072
2	30.0000	0.0285	0.065
			Average = 0.068

Uranium

By the emanation method.	
Average of two determinations	0.054%
Loss at 110°C. for three hours	0.012%
Loss on ignition	0.07%

If we assume the absence of ordinary lead and use the conversion factor 0.36 for thorium, the lead-uranium ratio is:

$$\frac{.068}{.054+3.44\times .36}=0.053.$$

The approximate[11] age of the mineral is:

$$\frac{.053\times 1.15}{1.57\times 10^{-4}}\text{million years}=388 \text{ million years.}$$

The age corresponding to a lead-uranium ratio of 0.053, reading from Holmes[12] graph is 391 million years.

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[10] The determinations by the emanation method were made with the assistance of Mr. Charles M. Burton.

[11] Kirsch, G., Geologie und Radioaktivität, **1928**, p. 128. The value 1.57 as given by Kirsch on p. 134, was used in the above formula, instead of 1.5.

Using the formula[13]:
Age in million years $= 15140 \log(1+1.155\text{Pb}/(\text{U}+.36\text{Th}))$ we get $t=391$ million years.
This would be the age of Taconic intrusions such as the Bedford cyrtolite.

The experimental result of this work is that the lead-uranium ratio of the Portland monazite is 0.053, and that the percentage of lead present in the mineral is small.

This research was greatly helped by a grant from the Elizabeth Thompson Science Fund and the writer expresses his grateful appreciation. The Laboratory of the University of Missouri and other privileges were made available to the author through the kindness of Dr. H. Schlundt.

Note by A. C. Lane

Originally two pounds and twelve ounces were sent by Spence, November 16, 1933. Eighty grams were specially purified by E. S. Larsen for O. B. Muench, of which 3 grams were sent to F. Hecht who made the pilot micro-analysis given below.

It will be noted that the lead is much less than Muench found. The difficulty of getting strictly reliable results on the lead in microanalysis, when it exists only in fractions of a per cent is shown by the lack of close agreement with Muench's figures. Yet it has value in that it confirms Muench's surprisingly low results. The thorium checks reasonably. It was hoped and expected that the mineral would have a ratio characteristic of the pre-Cambrian. It does not and there seems to be hardly enough lead for atomic weight determination. The question whether there can have been a loss of half the lead in the few years exposure in the dump, or whether there was some mineral rearrangement of this pegmatite at the time of the Taconic uplift requires further investigation. Looking at the material it is hard to believe a lead extraction. It seems fresh and the map shows not far away an igneous intrusion and it is conceivable that lead may have been driven off at the time of this intrusion.

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[12] Holmes, Arthur, on pages 208–9 of The Age of the Earth, Bulletin No. 80, *National Research Council*, June, 1931.

[13] *Report Com. on Measurement Geol. Time*, 1934.

Microanalysis of the West Portland Monazites (F. Hecht)[14]

Employed for analysis:	23.65 mg.	
SiO_2 (impure)	3.21%	
(SiO_2 pure)		(2.35)
PbO		
Pb		(.032 to .05)[1]
Rare earths[3]	62.09	
ThO_2	4.25	
(Th)		(3.72) or (3.18)[1]
U		(.094)[2]
Fe_2O_3	2.10	
Al_2O_3	0.69	
CaO	0.99	
MgO	0.56	
P_2O_5	27.39	
Loss on ignition (1000°)	0.60	
S	none	
Sum	101.88%	

[1] From later letters, the last letter (Apr. 7, 1935) reported that testing a larger quantity for Pb he has not been able to get more than 0.05 Pb, but that the Th checks better with Muench.

[2] 1.0035 grams was taken for this determination.

[3] The excess is due to the rare earths being weighed and reported as RO_2 which in the mineral are R_2O_3.

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[14] Methods and analysis reported by F. Hecht and E. Kroupa: *Zeit. für An. Chemie*, vol. **102**, pp. 81–90.

NOTES AND NEWS

CRYSTAL GROWTH AND SOLUTION UNDER LOCAL STRESS

George A. Russell, *University of Minnesota.*

It has often been stated that crystals dissolve where intensely strained and grow where the strain is less[1] but the idea that a schistosity may result from such oriented growth has been questioned. Riecke's principle,[2] based on his theoretical analysis indicates that a crystal under uniform stress is not in equilibrium with a solution that was saturated before the stress was applied: it will begin to dissolve. If another crystal in the solution is not stressed it will grow as the stressed crystal dissolves. Nevertheless, such an experiment and such a principle cannot be used to explain a development of elongated grains. It is emphatically stated that such an attempt would be based on a misunderstanding of the principle—it would be a misapplication of the idea.[3] This leaves some question whether such changes in relative length and thickness as occur during metamorphism can be attributed to solution under stress and resultant changes in index of elongation. A search of the literature revealed no definite experiment, although the suggestive statements are numerous.[4] This paper reports some simple experiments that can readily be repeated almost anywhere.

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[1] Harker, A., Metamorphism, pp. 145, 171, 1932.

Van Hise, C. R., Treatise on Metamorphism: *U. S. Geol. Survey, Mon.* **46,** p. 693. 1904.

Grout, F. F., Petrography and Petrology, pp. 398–400, 1932.

Taber, S., Growth of Crystals Under External Pressure: *Am. J. Sci.*, 4th Series, vol. **41,** pp. 532–556, 1916.

[2] Riecke, E., Zur Erniedrigung des Schmeltzpunktes durch einseitigen Zug oder Druck; *Centralbl. für Min., Geol., Pal.*, pp. 97–104, **1912.**

[3] Schwinner, R., Scherung, der Zentralbegriff der Tektonik: *Centralbl. für Min., Geol., Pal.*, pp. 474–475, **1924.**

Knopf, E. B., Petrotectonics: *Am. J. Sci.*, vol. **25,** pp. 460–461, June, 1933.

[4] Sorby, H. C., Devonian Limestones: *Proc. Quart. Jour. Geol. Soc.*, vol. **35,** pp. 87–89, 1879.

Harker, A., *op. cit.*

Wright, F. E., and Hostetter, J. C., The thermodynamic reversibility of the equilibrium relations between a strained solid and its liquid: *Jour. Wash. Acad. Sc.*, vol. **7,** pp. 405–417, 1917.

Wright, F. E., Schistosity by Crystallization: *Am. J. Sci.*, 4th Series, vol. **22,** pp. 224–230, July, 1906.

Van Hise, C. R., The pre-Cambrian rocks of the Black Hills: *Bull. Geol. Soc. Am.*, vol. **1,** p. 224, 1890.

THE OPTICAL DETERMINATION OF LOCAL STRAINS

(A) IN PYRALIN AND GLASS.

The phenomena termed photoelasticity were observed by Sir David Brewster in 1814. About a hundred years later the idea was applied by Groth,[5] and by Rinne[6] to determinations of the loci of stresses. Recently Coker[7] has applied the methods to metallurgical and engineering problems and Nadai[8] to geological problems. The interference colors of a stressed block of pyralin move progressively into the block from the points of application of stress. Figure 1-A shows two stages in the deformation of a pyralin block, the increase in stress being indicated by the large arrow.

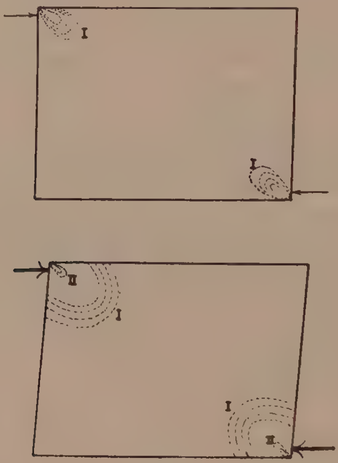

FIG. 1A

(B) IN CRYSTALS.

Rinne[9] showed by using garnet that the same localization of strain occurs in crystals. If stress is applied at a point or on a very small surface the interference color rings expand regularly with increasing stress. The writer agrees with these results in every way and can only offer a simplified verification that will be available in any petrographic laboratory. If a cleavage block of halite about $\frac{1}{2}\times\frac{1}{2}\times\frac{1}{8}$ inch is stressed by a small clamp (even a pinch cock for rubber tubing will serve) the colors can be developed on a petrographic microscope between crossed nicols. They can be watched spreading from the point of application as the screw is tightened, (Fig. 1-B). A small nail or wire serves to localize the pressure of the clamp.

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[5] Groth, P., Optical Properties of Crystals, pp. 261–282, **1910**.

[6] Rinne, F., Beitrag zur Kenntnis von Spannungdiagramen: *Centralbl. für Min., Geol., Pal.*, pp. 121–134 and 209–216, **1926**.

[7] Coker, E. G., *Metallurgia Magazine*, April, **1932**.

[8] Nadai, A., Plasticity, **1931**.

[9] Rinne, F., *op. cit.*

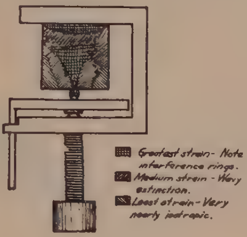
▦ Greatest strain- Note interference rings.
▨ Medium strain - Wavy extinction.
▧ Least strain - Very nearly isotropic.

FIG. 1B

It seemed perfectly evident that strains can be developed in a small part of the crystal (as in a small part of a glass or pyralin block) long before such strains reached all through the crystal. The strains may not be absolutely nil in the parts showing no interference color, but they must be very small.

SOLUTION EXPERIMENTS ON SINGLE CRYSTALS.

The direct attack on the problem was made with cleavage pieces of salt and crystals of ammonium alum, ranging from $\frac{1}{2}$ to 1 inch in size. A number of experiments were tried and while all tended to indicate the same principle, the very soluble alum gave the clearest demonstration. A measured, weighed, smooth-surfaced alum crystal in a saturated alum solution at constant temperature suffers no appreciable change for several weeks when evaporation of the solution is prevented. It was desired to find out whether a local strain in such a crystal would cause solution of the strained part and practically simultaneous growth of the other parts of the crystal.

A pressure rod was applied to the upper horizontal face of a crystal near one corner. The rod had a diameter of 1.5 mm. and an alum crystal will commonly stand a load of three or more kilograms on such a surface without crushing. Around the crystal was a solution saturated with alum at the temperature to be used. To facilitate the later determination of solution and growth, a little chrome alum was added. The growth from this solution was colored whereas the original was a nearly transparent white.

The temperature was kept constant within a degree Centigrade, by putting the container in a tank in which water was circulating under control by a toluene thermoregulator.

When all was set up the saturated solution around the crystal was covered with a layer of thick lubricating oil to prevent evaporation.

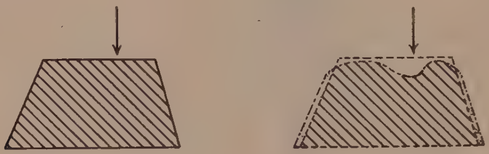

Fig. 2

In the most satisfactory experiment a solution and crystal were kept at 28°C. to 29°C. for 144 hours. The crystal which weighed 10.290 grams before, weighed 10.282 grams after treatment. It was corroded and remained white under and near the pressure rod, but had a growth of colored alum on all four sides in tiny parallel crystal faces. No other crystals had grown in the container. It is believed that the slight loss in weight (8 mgs.) may be a result of the pressure, or possibly of some imperfections in the apparatus. The volume lost from the top of the crystal was carefully estimated by restoring its original top dimensions with plasticene. This indicated the solution of over 0.6 gram from the top and a growth of nearly the same amount on the sides (Fig. 2).

Summary and Discussion

It seems perfectly clear that stresses may be localized in crystals as well as in amorphous material and that the solubility of a part of the crystal is related to the local stress.

Material dissolved from one part of a crystal because of a local stress may be redeposited on another part of the same crystal where the stress is relatively less, thus changing the form of the crystal.

This is perhaps not a statement of the Riecke principle nor an application of it, but it seems none the less to be a valid idea.

It is still too little known to be applied very far in metamorphism. The crystals in schists are known to be oriented crystallographically as well as elongated, and these experiments have not explained such an orientation. Possibly the change in shape here obtained may be perfectly possible under laboratory conditions, but if such conditions are rare in nature, the applications may not be important. The questions remain: (1) whether during mountain making movements the stresses applied to crystals are localized at certain corners as in the experiments here described, and (2) what other forces acting during metamorphism orient the crystal axes as well as the greater dimensions of the crystal.

PIEDMONTITE FROM LOS ANGELES COUNTY, CALIFORNIA

Russell R. Simonson, *University of California at Los Angeles.*

Petrographic studies of some of the metamorphic rocks of the Sierra Pelona schist series[1] have shown that piedmontite occurs as small euhedral crystals in quartz-sericite-biotite schist and in quartzite. The samples studied were obtained near the junction of Bouquet and Texas canyons in northern Los Angeles County.

Although manganiferous schists and quartzites are rather widespread in this region, the piedmontite is restricted to a zone about 300 feet wide and a half mile long adjacent to a quartz diorite dike which is three miles long and over 300 feet wide. The thermal effects and solutions from this intrusive aided the replacement of the laminae of biotite by piedmontite so commonly noted in the thin sections. Some of the quartzites show complete replacement; in such cases little or no biotite is found and parallel rows of pink crystals of piedmontite occupying former bands of biotite give the rocks a reddish cast.

Slides of several samples of the schist and quartzite showed that idiomorphic crystals of piedmontite less than 1 mm. long and 0.5 mm. wide were oriented parallel to the foliation; occasionally when an abnormal amount of the mineral is present radiating rosette-like masses are typical. Some of the crystals have been stretched and the interstitial cracks filled with secondary quartz while others nearby show no signs of deformation indicating several periods of piedmontite deposition. Pleochroic colors are: X = orange to lemon-yellow; Y = amethystine red; Z = carmine. A positive biaxial interference figure showed 2V to be large (70 to 80°). Extinction $X \wedge c$

is −3°; refraction and double refraction, high. The associated heavy minerals are manganomagnetite, psilomelane, garnet, titanite, and rutile.

As far as the writer knows, this is the first account of piedmontite from the meta-sedimentary rocks of California. Mayo described two specimens[2] and later a locality[3] of piedmontite in the metamorphic volcanic rocks of Madera County. Other California occurrences (in detrital material) are briefly summarized: Rogers[4] found piedmontite in a quartz porphyry boulder at Pacific Beach, San Diego County; A. S. Eakle[5] mentioned a piedmontite-bearing rock fragment from San Bernardino County; the mineral was reported from the lacustrine sediments of the Lizard area, southern San Joaquin Valley by Reed and Bailey,[6] O. A. Woodford[7] listed piedmontite as a constituent of the San Onofre breccia, and E. Wayne Galliher[8] found it in the bottom sediments of Monterey bay. A pebble of quartz-sericite schist from San Gabriel Canyon, Los Angeles County, found by W. J. Miller and examined by Joseph Murdoch[9] (both of the University of California at Los Angeles) contained piedmontite as an essential constituent.

It is probable that the distribution of this mineral in the Sierra Pelona series is more widespread than the present studies indicate.

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3. Mayo, E. B., Discovery of Piedmontite in the Sierra Nevada: *Calif. Dept., of Nat. Resources, Division of Mines*, Rept. of the State Min. for 1933, pp. 239–243, 1933.
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5. Eakle, A. S., Minerals of California: *Calif. State Mining Bur., Bull.* **67**, 1914.
6. Reed, R. D., and Bailey, J. P., Subsurface Correlation by Means of Heavy Minerals: *Bull. Am. Assoc. Petrol. Geol.*, vol. **11**, pp. 359–368, 1927.
7. Woodford, O. A., The San Onofre Breccia; Its Nature and Origin: *Calif. Univ., Dept. Geol., Bull.* **15**, pp. 159–280, 1925.
8. Personal communication to E. B. Mayo (see reference No. 3, p. 243).
9. Personal communication.

BOOK REVIEW

INTERNATIONAL TABLES FOR THE DETERMINATION OF CRYSTAL STRUCTURES. EDITED BY C. HERMANN, W. H. BRAGG, M. VON LAUE. Vol. I Tables on the Theories of Groups, XII+452 pages, 464 figs. Vol. II Mathematical and Physical Tables, VIII+240 pages, 83 figs. Gebrüder Borntraeger, *Berlin;* Chemical Catalogue Co., *New York.* **1935.** Price 33.—RM (paper); 40.—RM (bound).

This comprehensive work is the outgrowth of a conference of crystallographers and research workers in crystal structure in London in 1929. It forms an international standard reference work with a nomenclature to which all papers on crystal structure may be referred. The explanatory material preceding each chapter is repeated in German, English, and French; the major portion consists of tables and diagrams, with German context.

The scope and completeness of this work can be shown by a partial summary of the contents: Chapters I and II—A complete description of the Hermann-Mauguin system on nomenclature, with comparative tables of nomenclatures in common use. Throughout the text both Hermann-Mauguin and Schoenflies symbols are used. III—Survey of crystal classes. IV—Translation groups and transformations. V—Space groups. Each group is represented by two diagrams, one showing the distribution of symmetry elements, and the other the general positions of equivalent points. Coordinates of general and special positions are given, with the structure factor for the general positions and the laws of missing spectra. VI—Tables for space-group determination. VII—Point-symmetries. VIII—Lattice complexes. IX—Quadratic forms for $\sin^2 \theta$-values. X—Trigonometric and exponential functions. XI—Physical tables. A comprehensive article (in English) on intensity formulae; atomic form factors; absorption coefficients; ionic and atomic radii, etc. XII—Summary of graphical methods for evaluating diagrams and determining structures.

Through the cooperation of many workers, an enormous amount of data from widely scattered sources has been brought together in these two volumes. They represent a distinct contribution to the rapidly growing field of crystal structure. Not only will this compilation be of great convenience to structure workers, but it should put an end to the great variations which have existed with respect to nomenclatures, abbreviations, choice of axes and origins, etc.

The publication of these tables was made possible through grants made by ten organizations in the United States, England, France, Germany and Holland. The editors and authors served without remuneration.

L. S. RAMSDELL

NOTICE

The Secretary's Office wishes to notify the membership of the Society that through an error the date for the Annual Meeting was given in the September issue of the Ameican Mineralogist as December 27th-29th, 1935.

The Annual Meeting of the Mineralogical Society of America for 1935 will be held in the Waldorf-Astoria Hotel, New York City on

Thursday, December 26th
Friday, December 27th,
Saturday, December 28th.

PAUL F. KERR, *Secretary.*

NEW MINERAL NAMES

Repossite

GRILL, E.: Su un fosfato di ferro e di manganese delle pegmatiti di Ogliasca. *Period. Mineral.*, vol. **4**, No. 1, 19–23, 1935.

NAME: In memory of Prof. Emilio Repossi.

CHEMICAL PROPERTIES: A phosphate of iron and manganese: $(Fe, Mn)_3 (PO_4)_2$. Analysis: (by Paolo Gallitelli) P_2O_5 38.94, FeO 32.33, MnO 23.32, CaO 4.50, H_2O $-110°$ 0.15, H_2O+ 110° 0.39; total 99.63. Soluble in HNO_3 and dilute HCl. BB. fuses with effervescence to a black magnetic bead.

PHYSICAL AND OPTICAL PROPERTIES: Color salmon pink on fresh fracture, dark brown on exposure. Streak pink with tinge of yellow. Luster slightly greasy. Biaxial positive, optical angle 68°. $\alpha=1.708$, $\beta=1.713$, $\gamma=1.724$. Hd.$=4\frac{1}{2}$–5. Sp. Gr.$=3.72$–3.76.

OCCURRENCE: Found in pegmatite at Olgiasca, Lake Como, Italy, with quartz, feldspar, Muscovite and biotite, sometimes with colorless apatite and pyrite, and also pyrrhotite, arsenopyrite and vivianite.

W. F. H.

"Maxixe-aquamarine" Maxixeberyl

WILD, GEORG O.: Mitteilung über ein anscheinend neues Beryllium Silikat. *Centr Min.*, Abt. **A**, No. 1, pp. 38–39, 1933. SCHLOSSMACHER, K. AND KLANG, H.: Der Maxixeberyll. I. *Ibid.*, No. 2, pp. 37–44, 1935. ROEBLING, W. AND TROMNAU, H. W.: Maxixeberyll. II. *Ibid.*, No. 5, pp. 134–139, 1935. An alkali beryl (Li_2O 0.98, Na_2O 1.28, Cs_2O 2.80) with a small boron content (B_2O_3 0.39), of cobalt blue color and strong pleochroism has been called "Maxixe aquamarine." It is found at the Maxixe mine, south of Arassuahy, Minas Geraes, Brazil, associated with light rose tourmaline. The fine blue color fades rapidly upon exposure to sunlight to pale yellow or colorless. $\epsilon=1.58442$, $\omega=1.59203$.

W. F. F.

Aurosmirid

ZVIAGINCEV, O.: Ein neues metalle der Platingruppe enthaltendes Mineral. *Compt. Rend. Acad. Sci., U.R.S.S.*, vol. **4**, No. 3, 176–178 (Russian), 178–179 (German).

CHEMICAL PROPERTIES: Iridium, osmium and gold. ANALYSIS: Ir 51.7, OS 25.5, Ru 3.5, Au 19.3, Fe tr., Insoluble in aqua regia.

CRYSTALLOGRAPHICAL PROPERTIES: Cubic; structure cubic, face centered; $a=3.816$Å.

PHYSICAL AND OPTICAL PROPERTIES: Color pale silver white. Luster metallic. Slightly malleable to brittle; fracture uneven; no cleavage. Hd.>7. Sp. Gr.$=20$.

OCCURRENCE: Found as irregular grains 0.5–2 mm. in size in the residue insoluble in aquaregia of platinum; associated with crystals of osmiridium.

W. F. F.

Mr. Paul Hines of the University of Virginia has been appointed Teaching Fellow of Mineralogy at Stanford University for the year 1935–36.